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LINEATION, SYMMETRY, AND INTERNAL MOVEMENT IN MONOCLINIC TECTONITE FABRICS

BY FRANCIS J. TURNER

ABSTRACT

The current controversy regarding kinematic interpretation of lineation in schists hinges on the significance of symmetry in tectonite fabrics. Interpretation of such fabrics by Sander and Schmidt is based on an assumption that symmetry of fabric reflects symmetry of internal movements accompanying deformation. In identifying *b* lineation, which is the symmetry axis of certain monoclinic fabrics, as the principal direction of movements concerned in the evolution of those fabrics, some recent writers have neglected or rejected Sander's postulate regarding symmetry. This course seems unjustified for two reasons: (1) Those who identify regional lineation with movement direction have done so for Precambrian or Paleozoic rocks whose metamorphic and deformational history is ambiguous and often highly complex; the "direction of movement" in such rocks has not been established independently of fabric evidence. (2) Since it was proposed 30 years ago, the symmetry concept of Sander and Schmidt has become strengthened by evidence accruing from studies on fabrics of experimentally deformed metallic aggregates and ceramic bodies. Additional supporting evidence now comes from fabrics of experimentally deformed marble. In deformed marble cylinders symmetry of fabric is identical with symmetry of movement as inferred from measurable strain. The pattern of strain is controlled by the orientation of the cylinder (a mechanically anisotropic aggregate of grains) in relation to applied force. Even in highly deformed material (*e.g.*, where elongation exceeds 500 per cent) the influence of the original anisotropy on symmetry of experimentally induced movement and strain is obvious in the final fabric.

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Acknowledgments

Our knowledge of the fabric of experimentally deformed rocks is based almost entirely on specimens of marble, dolomite rock, granite, and dunite deformed at 20–500° C and 3000–5000 atmospheres by Professor D. T. Griggs and associates in the Institute of Geophysics, University of California, Los Angeles. Both the experimental program of Professor Griggs and subsequent petrofabric analysis of the deformed material by the writer and associates have been generously supported by grants from the Penrose Bequest of The Geological Society of America, the Office of Naval Research, U. S. Department of the Navy, and the National Science Foundation. To the John Simon Guggenheim Memorial Foundation I am indebted for the opportunity to see at first hand the Moine and Dalradian schists of the Scottish Highlands, and to discuss with Professor Sander and general subject of this paper. I wish to thank Mrs. E. B. Knopf for critical reading of the manuscript, and Miss B. R. Dixon who drafted the figures.

Kinematic Significance of Symmetry in Analysis of Tectonite Fabrics

Bruno Sander's retirement from the chair of mineralogy at the University of Innsbruck marks the close of an era in the development of modern structural petrology (*Gefügekunde*). Largely through the work of the Innsbruck school over more than 4 decades the descriptive side of structural petrology has been placed on a firm basis. The statistical methods of petrofabric analysis have been applied with equal success to microscopic features (notably those relating to preferred orientation of mineral grains) and to megascopic elements in the fabric of deformed rocks (bedding, foliation, lineation, jointing). Common patterns of preferred orientation of quartz, mica, calcite, and other minerals metamorphically deformed are now well known. Petrogenically they are comparable in importance to the accumulated data relating to mineral facies. Also well established by Sander and coworkers is the uniformity of symmetry commonly displayed, within a given mass of deformed rocks, by fabric elements ranging from microscopic characters to large-scale features the nature of which can be demonstrated only by mapping in the field.

Kinematic and dynamic interpretation of tectonite fabrics is conspicuous in the publications of the Innsbruck school. This aspect of structural petrology necessarily rests on a less secure foundation than the descriptive phase. It contains much that is speculative; and I would now reject as unwarranted many published conclusions, reached in Austria and elsewhere, with regard to hypothetical processes of grain orienting. This applies especially to interpretation of preferred orientation patterns—for example those of quartz—in terms of hypothetical mechanisms of crystal gliding unsupported by experimental data. Nevertheless, what is perhaps the greatest achievement of Sander and his former coworker Schmidt lies within the controversial field of interpretation. From the outset, when he published his classic paper (Sander, 1911) "On the relation between rock fabric and movement of component parts", Sander has favored kinematic rather than dynamic interpretation of rock fabrics. To correlate fabric with internal movements is less doubtful than is more tenuous correlation with forces responsible for such movements.

Symmetry has emerged as the basic criterion for correlating fabric with movement (Cf. Schmidt, 1926; Sander, 1930, p. 53–73, 145–147; 1948, p. 66–83; Knopf and Ingerson, 1938, p. 42–62). On the basis of experience drawn from such fields as metallurgy and hydraulics it was assumed that the symmetry of a tectonite fabric reflects the symmetry of movement responsible for the evolution of the fabric. Analogies were drawn with familiar examples of preferred orientation of moving objects—wheat stalks bent in the wind, logs in a flowing stream, sand in dunes, birds in flight. Today the symmetry rule stands as a well-tested postulate supported by an imposing body of experimental evidence drawn from such diverse fields as hydrodynamics, sedimentation, metallurgy, and ceramics. Also significant is the persistence of fabric symmetry throughout successive phases of metamorphism. Although crystallization commonly outlasts deformation in metamorphism, it seldom obliterates, and indeed often emphasizes symmetry of fabric imprinted by deformation. In using symmetry to interpret

rock fabrics two points should be borne in mind: the over-all plan of movement deduced from fabric criteria applies only to the field within which the fabric has been investigated (a thin section, hand specimen, single outcrop, or map area); and the validity of the symmetry rule is independent of assumptions and inferences (commonly highly speculative and unwarranted) regarding the mechanisms of rock deformation or of grain orienting.

Symmetry and Lineation in Monoclinic Movements and Fabrics

Monoclinic symmetry of movement is illustrated by flow of water under gravity in a flat-bottomed channel of symmetrical cross section. The vertical plane containing the direction of flow is the sole symmetry plane; and normal to it is a unique axis of binary symmetry. Where flow is laminar, sense of flow is the only feature that identifies the symmetry of movement as monoclinic rather than orthorhombic (with three planes and three binary axes of symmetry). For turbulent flow symmetry is more obviously monoclinic. Here the sole binary axis is further emphasized by rotational movements, fluctuating in direction, space, and time, around axes whose mean trend is the symmetry axis of the integrated movement plan. Where flow occurs in a channel the width of which symmetrically increases downstream, the dominant downstream flow is accompanied by lateral spread in both senses parallel to the symmetry axis. But symmetry of the whole movement plan is still monoclinic.

Flow may be described with reference to three mutually perpendicular axes *a*, *b*, and *c* selected (as in crystallography) so as to bring out as simply as possible the geometric properties of the system. In the monoclinic movement plan *a* is defined as the direction of flow and *b* as the normal to the single plane of symmetry. Where flow is laminar *a* is the most significant of the three axes; but where flow is turbulent *b* is the most consistent of the three axes and so dominates the movement plan.[1] It is a matter of common observation that rods and plates suspended in the flowing stream assume a state of preferred orientation with their longer dimensions parallel either to *a* or to *b*. Moreover this state tends to be perserved in deposited sediment after flow has ceased. The pattern that develops depends upon such factors as particle dimensions, density of particles and of fluid, viscosity, and rate of flow; but its symmetry invariably is monoclinic. Orientation of sand dunes, regular patterns of minor ripples on their surfaces, and the laminated fabric and sorting of the dune sand have in common a monoclinic symmetry that faithfully reflects symmetry of observed wind movement.

In fabrics of metamorphically deformed rocks monoclinic symmetry likewise is common. By analogy with the monoclinic movement plan, the single symmetry plane of such a fabric is identified as *ac*. The single symmetry axis *b* dominates many monoclinic fabrics by virtue of large- and small-scale folding and other effects of external rotation around *b*. But in some monoclinic fabrics, especially those of mylonites and slickenside films, the direction of transport *a* is the dominant axis, external rotation around *b* is insignificant, and the fabric may approach orthorhombic symmetry.

The most consistent and conspicuous element of many metamorphic fabrics having monoclinic symmetry is lineation. It is defined by such characters as axes of microfolds, intersection of *s*-surfaces, parallel alignment of tabular or prismatic crystals, elongation of lenticular or rodlike aggregates of mineral grains, and parallel grooving of surfaces of shear (as in slickensides). Lineation may coincide with either *a* or *b* of a simple monoclinic fabric. Interpreted according to the symmetry rule of Sander and Schmidt, *a* lineation represents a direction of differential movement or shear, whereas *b* lineation is transverse to the mean direction of tectonic flow. In tectonic interpretation of lineated fabric it is clearly of paramount importance to distinguish between these two. The distinction is based on symmetry patterns first recognizable in the fabric as a whole and then correlated, according to Sander's rule, with symmetry of the movement plan of the deformation.

The above discussion is confined to common simple monoclinic tectonite fabrics. Interpreta-

[1] For comparison of flow in rocks with laminar and turbulent flow of fluids, *see* G. Wilson (1946, p. 296–297; also in discussion of a paper by Anderson, 1948, p. 127).

tion of more complex fabrics resulting from repeated deformation or possessing triclinic symmetry is beset with ambiguity that cannot be resolved without additional information.

Critical Examination of Attempts to Correlate *b* Lineation With *a* of the Movement Plan

General Statement

Opposition to kinematic interpretation of strongly lineated fabrics in terms of the Sander-Schmidt symmetry principle is apparent in several recent papers on structural petrology: Strand (1944); Kvale (1945; 1948; 1953); and Oftedahl (1948) in Norway; Anderson (1948) with reference to the Moine schists of Scotland; and Balk (1952; 1953) and Johnston (1954) in North America. These writers cannot reconcile the movement plan deduced from fabric symmetry with that inferred from regional geological mapping. They have resolved their dilemma by rejecting the symmetry concept; and they conclude that the *b* axis of monoclinic fabrics, even where it is the axis of small-scale recumbent folding, is commonly parallel to the *a* axis of movement. In addition Kvale (1948, p. 247; 1953, p. 57, 60) notes the occurrence of complex fabrics with several intersecting lineations, and concludes from this that trend of lineation need bear no simple relationship to principal direct of movement. These radical departures from thodox interpretation of rock fabrics based marily on symmetry merit careful scrutiny.

Movement in Regional Deformation

Underlying the argument leading to rejection of the symmetry principle are two basic assumptions—either implicit or explicitly stated: (1) Deformation in any major sector of a fold belt (orogen) is the result of movement in one principal direction; and movement in any other direction is of minor significance. (2) Mapping on a rather broad scale establishes the principal movement direction: it is parallel to the direction of crustal shortening, and normal to the local trend of the folded zone and to the strike of the folded rocks.

To me this picture of movement in orogeny seems much too simple. Admittedly recumbent folding and thrusting afford convincing evidence of one component of movement—horizontal or inclined—normal to the general trend of fold arcs. Such movement within the deformed mass is possible only in the upper levels where the material so affected is squeezed out from between the mutually approaching walls of the collapsing trough. However, there is no reason to believe that this is the sole or even the predominant component of movement in all orogeny. In spite of contrary statements (*e.g.*, Kvale, 1945, p. 22), a component of movement acting parallel to the axis of regional folding is likely to be effective. If, in conformity with today's fashion, subcrustal convection is invoked as the driving force of orogeny, the tectonic stream might commonly be directed obliquely to the trend of the zone of crustal weakness (orogen) in which deformation is localized. In upper levels the deformed material is free to spill upward and outward; it yields by recumbent folding and thrusting across horizontal or gently plunging axes, with the principal direction of movement transverse to the trend of the deformed zone. But at deeper levels movement is constrained to directions more nearly parallel to the length of the orogen. Here transverse folds and related movements across axes that commonly plunge steeply characterize the movement plan. Extensive lateral movement in the same general direction is now recognized in major late dislocations trending parallel to folded geosynclinal belts—*e.g.*, the Great Gelen Fault of Scotland and the San Andreas fault system in California.

The above tentative simplified picture illustrates my opinion that a complex movement picture of orogeny is more plausible than a simple picture involving one principal direction of movement. It is consistent with the remarks of Professor R. M. Shackleton in discussion following a recent paper of Kvale (1953, p. 67): "The speaker [Professor Shackleton] was inclined to think that with increasing importance coming to be attached to diapirism and intrusion tectonics in depth, and to gravity-operated flow near the surface, the assumption of rather uniform direction of transport was unjustified. It seemed that in the Scottish Highlands flow of plastic materials parallel to the

general trend of the orogeny might account for much of the transverse structures."

Critical Review of Field Evidence as to Movement Directions

The field evidence and underlying implicit assumptions on the basis of which the authors quoted deduced a single principal direction of movement is open to serious criticism:

(1) The regional trend of the fold arcs has been said (*e.g.*, Kvale, 1953, p. 61) to indicate principal movement at right angles to it. While generally applicable to upper levels where recumbent folds and nappes were free to develop, this inference cannot be extrapolated to deeper zones of high-grade metamorphism as typified by the Moine and Dalradian schists of Scotland.

(2) Much of the ambiguity that obscures interpretation of lineation in the Scottish Highlands has resulted from identifying the general trend of outcropping formations with the axis of folding (Cf. McIntyre, 1951b; 1952, p. 259).[2] Such identification is true only for folding about horizontal axes. Where axes of recumbent folds plunge noticeably, the general trend of outcrop is likely to be normal rather than parallel to fold axes (McIntyre, 1950, p. 428; 1952).

(3) The validity of correlating regional developed lineation with supposed directoin of localized movement in major thrusts is doubtful:

(a) Displacement parallel to the dip of a thrust is commonly so spectacular that the possibility of even more extensive displacement parallel to the strike tends to be overlooked. This situation is exemplified by the century-long controversy—still unresolved—regarding direction and age of movement on the Moine Thrust and associated dislocations in northwestern Scotland (*see* summary by McIntyre, 1954). In the absence of detailed mapping, the conclusions of Kvale (*e.g.*, 1948; 1953, p. 57) regarding movement on thrusts in the Bergsdalen area of Norway, and of Balk (1952, p. 416–419) with reference to direction of thrusting in Dutchess County, New York, are opinions rather than established facts. This applies also to Johnston's (1954, p. 1059) assessment of movement of Precambrian thrusts in Canada. Balk (1953, *e.g.*, p. 839) assumed that the direction of movement within plates of highly deformed rock occurring in thrust zones in the Taconic area of New York must have coincided with the dip, even though the assumed movement direction is also the axis of contemporary folding of the deformed material (Balk, 1953, Pl. 6, fig. 3). In the evidence he presented, there is nothing to justify this assumption, on the basis of which Balk identified the *b* axis of monoclinic fabrics as *a* of the movement plan.

(b) Progressive strengthening of regional lineation with increasing proximity to thrusts has been cited repeatedly as evidence connecting lineation with thrusting. In the absence of quantitative data this too is a matter of opinion, as is apparent in published discussion on possible connection between movement on the Moine Thrust and the development of mullion structure and other forms of *b* lineation in the Moine schists (Cf. G. Wilson, 1953; McIntyre, 1954).

(c) Mylonites and phyllonites in zones of thrusting commonly show intersecting lineations and other indications of repeated movement in several directions. Again the Moine Thrust may be cited as an illustration (Christie, McIntyre, and Weiss, 1954, p. 220).

(4) In every region where regional *b* lineation has been identified with the *a* direction of movement there is convincing evidence of repeated movements sometimes differing greatly in age. Thus Kvale (1945, p. 212) assumes complete obliteration of a Precambrian metamorphic fabric by Caledonian deformation in order to correlate the existing lineation with the *a* direction of Caledonian movement. Weiss (1953) has shown that in west Spitzbergen folding on a north-south axis (emphatically attributed to Caledonian deformation by Kvale, 1953, p. 61) has strongly affected Carboniferous sediments. It was preceded by deformation with folding (presumably Caledonian) on east-west axes. Johnston (1954, p. 1047) correlates lineation in the Grenville of Canada with the latest of a series of many Precambrian movements

[2] Current differences in interpretation of lineation in Central Australian granulites seem to stem largely from the transverse (north-south) trend of regional *b* lineation, with respect to the prevalent east-west strike of the metamorphic rocks and east-west trend of existing mountain ranges (A. F. Wilson, 1953.)

localized in fault zones. In Dutchess County, New York, where Balk (1952) attributed development of lineation to a simple thrusting movement, detailed mapping and structural analysis have revealed three successive deformations, each characterized by its peculiar movement plan (Knopf, 1954). In the current controversy regarding significance of lineation in Central Australian granulites, A. F. Wilson (1953) has shown that the regional lineation of these rocks is parallel to *b* of the fabric, and draws attention to the fallacy of correlating it with a supposed *a* direction of movement on thrusts which postdate the main regional metamorphism.

Critical Review of Evidence as to Movement on a Small Scale

Evidence of minor folds.—Where initially planar structures have been folded on a small scale, *e.g.*, within the field of a hand specimen or an outcrop, the deformed structure can be measured and the nature and direction of movements can be determined beyond doubt. The value of such information in studies of regional tectonics has long been recognized (Cf. Bailey, 1935, p. 51–52; Wilson, 1946, p. 291–292; 1953, p. 139–140). The movement plan of a minor fold holds good only within the field investigated. Nevertheless where the same plan is exhibited in outcrop after outcrop over a large province it acquires regional significance. Reasoning partly on these lines McIntyre (1951b), Wilson (1953, p. 140), and others are justified in concluding that deformation (with small-scale recumbent folding) in the Moine schists involved regional movement in a northeast-southwest or east-west direction. The above discussion pertains only to the common case where simple shear- or flexural-slip folds have monoclinic symmetry. Minor folding of a more complex type, often with triclinic symmetry, is also common where initially anisotropic rocks have been deformed, or in cases of repeated deformation (Cf. Weiss, 1955).

In most areas where regional lineation has been identified as the *a* direction of movement, minor folding about axes parallel to the lineation is nevertheless widely developed (*e.g.*, Kvale, 1948, Pl. 15; 1953, p. 55; Balk, 1952, p. 418; Johnston, 1954, p. 1059). In some such areas (*e.g.*, McIntyre, 1951a) a consistent sense of overturning in areas of many square miles emphasizes the regional uniformity of the monoclinic movement plan. Some advocates of *a* lineation dismiss the movement recorded in such folds as of minor significance. Others (*e.g.*, Anderson, 1948, p. 124–125) attribute minor folds to rotation around the *a* axis of movement. To me neither course seems justified.

Evidence of preferred orientation of quartz and mica in schist and in quartzite.—The literature on lineation in deformed rocks abounds in orientation diagrams for quartz and mica (*e.g.*, Phillips, 1937; Kvale, 1945; Balk, 1952; Johnston, 1954). Since virtually nothing is known as to the mechanisms by which these minerals become oriented during metamorphism, the only property of the diagrams that could now be of value for kinematic interpretation is their symmetry. This should be considered as one component in the over-all symmetry of the rock fabric. From this standpoint three symmetry types can be recognized in the fabrics referred to above:

(1) Orthorhombic: two symmetry axes, *a* and *b*, lie in the foliation; one of these coincides with the lineation which is also the axis of incomplete girdles for *c* in quartz and for {001} in micas; paired maxima are present in the quartz girdle (Fig. 1A). Symmetry is compatible with movement parallel or normal to the lineation. This is a rather rare type exemplified by Phillips (1937, Figs. D 1, 2, 3, 22).

(2) Monoclinic: a single symmetry axis, *b*, coincides with the lineation which is also the axis of girdles for *c* in quartz and for {001} in micas (Fig. 1B–D). Symmetry is compatible only with movements normal to the *b* lineation. This type of fabric is very common (Kvale, 1945, p. 201, Figs. 6–10; Phillips, 1937, Figs. D 6–14, 20, 21, 23, 24; Balk, 1952, Figs. 16, 24, 27; Johnston, 1954, D 38–43).

(3) Triclinic: axes of quartz or mica girdles are obliquely inclined to lineation and in some fabrics to the foliation (Fig. 1E). The symmetry is compatible with compound movements in two or more superposed unrelated deformations, or with deformation of an initially anisotropic fabric (Weiss, 1955); for examples see Balk 1952, Figs. 6, 10, 12, 13, 19, 21, 26).

Preferred orientation data collectively support the thesis that movement normal to the

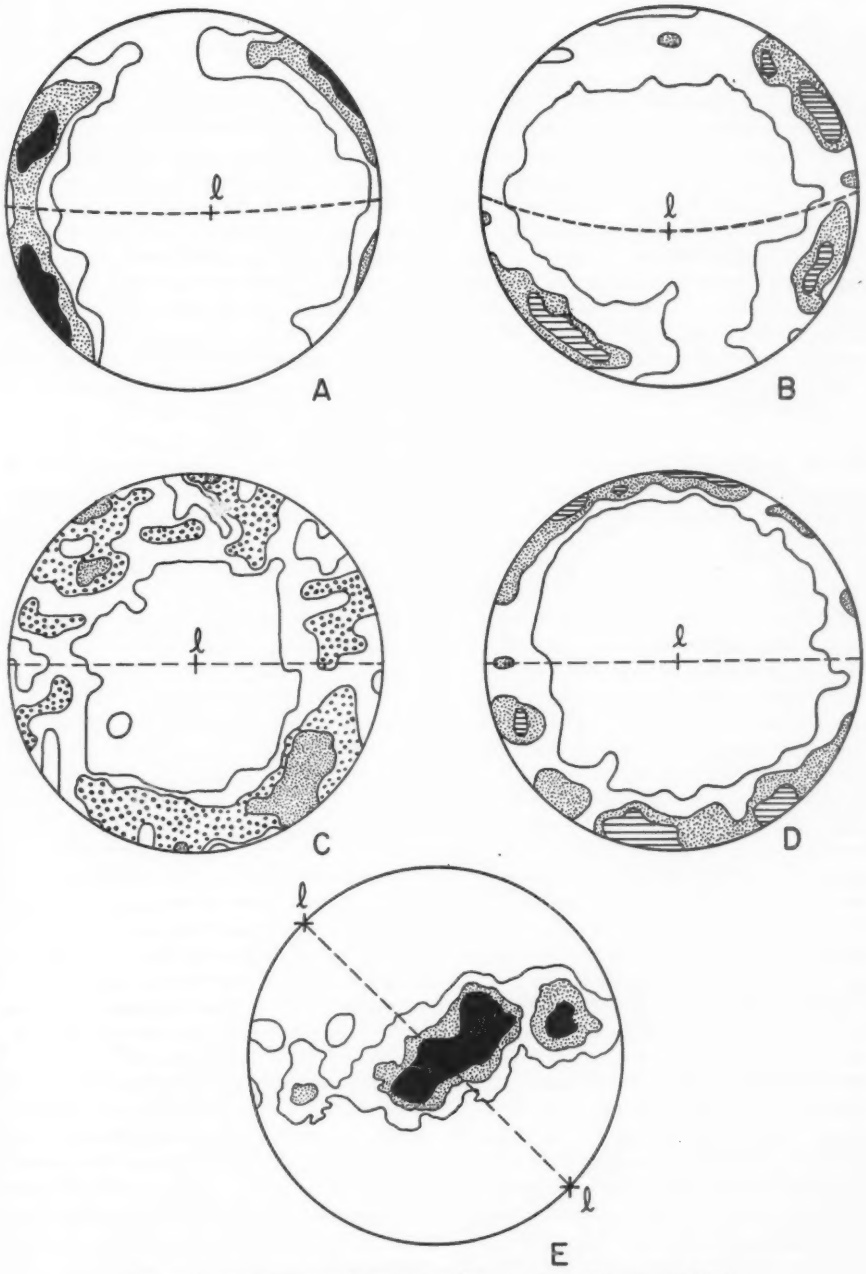

FIGURE 1.—SYMMETRY IN ORIENTATION DIAGRAMS FOR QUARTZ AND MICA

A–D. Moine schists (Phillips, 1937, D 22, 8, 23, 24). Broken arc = foliation; *l* = lineation. A. Quartz, 200 *c* axes. Contours 1, 3, 8 per cent. *Orthorhombic.* B. Quartz, 250 *c* axes. Contours 1, 3, 4 per cent. *Monoclinic.* C. Quartz, 250 *c* axes. Contours 1, 2, 3 per cent. *Monoclinic.* D. Mica, 200 normals to {001} in same rock as C. Contours 1, 3, 5 per cent. *Monoclinic.*

E. Quartzite, Dutchess County, New York (Balk, 1952, Fig. 21). Quartz, 300 *c* axes. Contours 1, 3, 10 per cent. *l* = lineation; broken line = trace of *s*-plane. *Triclinic.*

trend of regional lineation is responsible for the fabrics of the quartzites and schists described in the publications cited above. Repeated deformation with different movement plans is indicated by some fabrics, notably in the rocks from New York described by Balk. If fabric symmetry is abandoned as a criterion of movement, as advocated by Kvale and Anderson, then the preferred orientation patterns of quartz and mica yield no intelligible evidence bearing on the lineation problem.

Evidence of shape of deformed pebbles.—Initially spheroidal pebbles of conglomerates commonly survive severe deformation. Shape, internal fabric, and preferred orientation of quartzite pebbles in metaconglomerates have been carefully studied by Strand (1944), Oftedahl (1948), Flinn (1952), Brace (1955), and Elwell (1955), all of whom have used the data so obtained to reconstruct movement plans of deformation. The form of each pebble may be regarded as a strain ellipsoid—typically triaxial and highly elongate in the A direction.[3] It describes the strain within the field of an individual pebble, and is the tangible record of internal movements for the most part directed at right angles to the B axis.

Even if deformation within each pebble were purely mechanical, any of several distinct internal movement plans could account for the present form of the ellipsoid. Three simple possibilities are: (1) shear on one set of *s* planes which must be inclined to the AB plane but parallel to the B axis; (2) equal shear on two sets of *s* planes symmetrically inclined to AB and intersecting in B; (3) simultaneous shear on two pairs of *s* planes, one intersecting in B and the other in A. In all three symmetry of strain is orthorhombic, A, B, and C are axes of binary symmetry, and movement is solely or principally in directions transverse to B. There is nothing in the shape of the pebble to indicate preference for any one of the three movement plans. Moreover, yet other types of movement, *e.g.*, upon conical rather than plane surfaces of shear, are possible. Another complicating factor is the unknown role of recrystallization in relation of deformation. Ambiguity may be reduced, though not necessarily eliminated, by analysis of (1) internal fabric of pebbles and (2) megascopic fabric of the conglomerate as a whole. This has been attempted by Flinn (1952) and by Elwell (1955).

Evidence of internal fabric of deformed pebbles.—In the absence of experimental data on gliding and orienting of quartz or mica under metamorphic conditions, discussion of internal fabric in terms of imaginary orienting processes, as has been attempted by Brace (1955, p. 140–143) is so speculative as to increase rather than to diminish ambiguity regarding the general movement plan. Symmetry of preferred orientation in relation to other fabric elements remains the sole criterion upon which to develop a movement plan. On this basis the data on quartz orientation given by Strand (1944, p. 28, 29) and Brace (1955, p. 138, 139) are here reinterpreted as follows:

(1) Most of the diagrams for quartz axes have monoclinic symmetry which rarely approaches orthorhombic (*e.g.*, Strand, 1944, Figs. D 3, 5). Girdle patterns are characteristic. The sole (in orthorhombic diagrams the principal) symmetry axis coincides with the girdle axis and with the longest diameter A of the pebble. These features are consistent with an essentially monoclinic movement plan for each pebble, the *b* axis of movement being the longest diameter of the pebble. The Irish conglomerates described by Elwell (1955) likewise contain stretched pebbles whose long axes coincide with the axes of sharply monoclinic quartz girdles. Here the conglomerate fabric as a whole is monoclinic, and its *b* axis is defined by alignment of quartz pebbles, megascopic lineation, and axes of major and minor folds.

(2) Brace (1955, p. 141) notes that closely adjacent pebbles yield quartz-axis diagrams differing markedly from one another in detail, and attributes such differences to the influence of initial fabric prior to deformation. This is a logical inference. It emphasizes local departures from a regional movement plan. So also does the triclinic symmetry of some of Brace's diagrams (Brace, 1955, p. 139, nos. 7a, 9a).

(3) Strand gives one quartz-axis diagram (Strand, 1945, Fig. D 1) of radically different pattern. There is a strong girdle whose axis coincides with the intermediate diameter B

[3] The longest, intermediate, and shortest principal axes of the ellipsoid are designated A, B, and C respectively.

(instead of the long diameter A) of the pebble. Significantly this is the only diagram representing a pebble from the mylonitized zone on Strand's map. On the basis of symmetry several movement plans are possible. Strand prefers one involving movements at right angles to the B-girdle axis. I share this preference but, unlike Strand, would attribute the unique pattern of this single quartz fabric to local movements connected with mylonitization, the *b* axis of which was perpendicular to the *b* axis of earlier much more general movements responsible for the fabric of pebbles outside the mylonite zone. By contrast, Strand correlates every feature of the conglomerate fabric, whether local or regional, with simple shear parallel to the direction of regional lineation and pebble elongation.

Evidence of preferred orientation of pebbles.—Finally we must consider evidence provided by the state of preferred orientation of elongate pebbles in relation to other elements of the megascopic fabric. Correlation of the mean direction of pebble alignment with the *a* axis of a movement plan, which also takes into account the schistosity and lineation of the conglomerate raises major difficulties (Cf. Strand, 1944, p. 24; Oftedahl, 1948, p. 483; Brace, 1955, p. 133).

Strand and Oftedahl interpret the regional schistosity as shear surfaces, and the regional lineation (the direction of pebble alignment) as the direction of shear. But they realize that this simple movement plan, applied to individual pebbles as units, cannot explain the shape and present orientation of the pebbles. Oftedahl therefore postulates (1) internal shear on planes initially coinciding with the regional shear schistosity (Fig. 2A), and (2) subsequent external rotation of each pebble to bring its AB plane into the plane of schistosity (Fig. 2B). Such a rotation mechanism is purely imaginary, and there is no record of it in the fabric of the conglomerate matrix.

Brace (1955, p. 132, 133) in a similar dilemma deduces an even more complex movement plan. The principal schistosity of the conglomerate-schist complex is attributed to shear, and a fold lineation within it is identified as a *b* axis of movement (*l* in Fig. 3). This part of the picture seems well substantiated. Within the schistosity plane, and inclined at 60° to this *b* lineation, is the direction of alignment of elongate pebbles (*l'* in Fig. 3). The total symmetry clearly is triclinic and to me suggests superposed unrelated deformations. Brace, however, attributes the whole fabric to one compound deformation

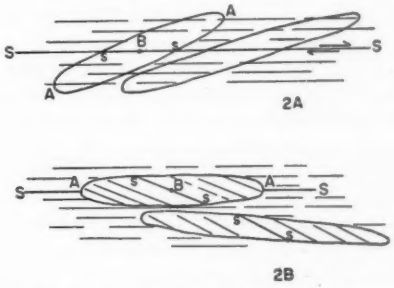

FIGURE 2.—SHEAR PLANES IN QUARTZITE PEBBLES (*ss*) AND ENCLOSING MATRIX (*SS*)]

As described by Oftedahl (1948). A. Simultaneous shear on *SS* and *ss*. B. Present orientation of pebble with AB plane approximately parallel to SS of matrix

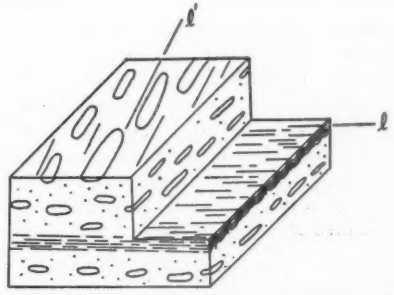

FIGURE 3.—RELATION OF PEBBLE ELONGATION (*l'*) TO LINEATION (*l*) IN METACONGLOMERATE

As described by Brace (1955). Schist band with lineation *l* and strain-slip is interbedded with pebble bands.

involving two totally different movement plans: (1) shear and crenulation localized in the schist layers, movement being normal to the present *b* lineation; (2) flattening movements within the conglomerate, the pebbles of which became elongated at 60° to *b* of movement plan (1). The problem has been clearly presented and discussed by Brace; but his solution seems improbable in that it involves simultaneous operation of movements which have no symmetry in common.

The main value of the preferred alignment of deformed pebbles is that it constitutes one of the elements contributing to the total symmetry of

the fabric. In the rocks described by Brace the direction of alignment of pebbles is inclined at 30° to the symmetry plane of an otherwise monoclinic folded fabric (Fig. 3). The over-all symmetry thus is triclinic. It is consistent with a compound movement plan involving two superposed unrelated deformations.

Flinn's (1952) tectonic analysis of deformed conglomerates in the Shetland Islands and Elwell's (1955) account of a deformed boulder bed in the Dalradian of Ireland differ from the work discussed above in two respects. The deformed pebbles are treated as but one element of a compound fabric; and an over-all fabric picture is synthesized by analyzing and comparing such diverse elements as the stretched pebbles, quartz orientation patterns in pebbles and in enclosing matrix, attitudes of foliation and lineation, minor fold axes, and major fold axes. In each the complete analysis reveals a simple monoclinic symmetry, with a strongly defined *b* axis parallel to which the elongated pebbles of the conglomerates are aligned. In the Shetland example, *b* is defined by megascopic lineation, *ac* jointing, and a sharp girdle of foliation surfaces. In the area described by Elwell, *b* is defined by megascopic lineation, minor and major fold axes, and quartz girdles in both conglomerate and matrix. In conformity with Sander's symmetry principle, both writers correlate *b* of the fabric with *b* of the movement plan.

Concluding Statement

Correlation of *b* lineation in monoclinic fabrics with the *a* direction of movement is here criticized on the following grounds:

(1) It negates the Sander-Schmidt concept of the kinematic significance of symmetry in deformed fabrics. This concept of course is not an established scientific law. But it is a reasonable hypothesis, well substantiated by observations in several fields of science.

(2) It assumes that a single direction of movement (*a* of the movement plan) within the whole mass of deformed rocks constituting a major fold-mountain belt, such as the Caledonian folds of Europe, can be read from a regional geological map. And this direction can then be applied universally "with absolute certainty" (Kvale, 1945, p. 204) to any minor portion of the total mass—a small map area, a single fold, a hand specimen, or a pebble in metaconglomerate. It applies likewise to all intervals of time within the orogenic period. To most students of flow and deformation such uniformity of movement in space and time would seem highly improbable. Much preferable is a model of turbulent flow. Petrofabric analysis of microscopic and megascopic features establishes symmetry of fabric within fields ranging in size from a hand specimen to an area of a few square miles. This reflects symmetry of movement within each of the analyzed fields. On such a basis may be built a compound movement plan showing how movement has varied from place to place, or even from time to time, in a single orogeny. An incomplete but already obviously complex regional movement plan for metamorphic deformation of the Moine schists of Scotland has already emerged from mapping of *b* lineation, begun by Phillips (1937) and continued by McIntyre (*e.g.*, 1951b), Sutton and Watson (1954), and others. The conflict between the complicated plan thus deduced and the simple plan of regionally uniform movement based on the broadest features of the geological map is no reason for discarding symmetry of fabric as a criterion of movement. Rather it emphasizes the value of petrofabric studies as the means for revealing otherwise unsuspected complexities in the kinematics of orogeny.

(3) Underlying correlation of the direction of pebble elongation with the *a* axis of movement is the assumption that great elongation of small elements in the fabric (*e.g.*, pebbles) can occur only in directions transverse to the *b* axis of the movement plan. Yet both Kvale (1945, p. 198) and Balk (1952, p. 426, 427) record considerable elongation of quartz grains at right angles to the main lineation which they identify as *a*. Elongation of different elements of a single fabric in two mutually perpendicular directions is, in fact, a well-known characteristic of some highly deformed rocks. Moreover at times (*e.g.*, Flinn, 1952; Elwell, 1955) complete tectonic analysis of deformed conglomerates in relation to associated rocks leads to the conclusion that the *b* fabric axis (direction of pebble elongation) must be the *b* axis of the movement plan. I attach particular significance to the behavior of heterogeneous rocks composed of small masses of stronger material, such as quartz,

separated by a laminar matrix of weaker material, such as mica. Flow within the latter may be the principal means of deformation. The stronger bodies are stressed somewhat as is a marble cylinder during an elongation experiment at high confining pressure (Cf. Griggs and Miller, 1951, p. 856). The compressive stress acting in any direction normal to the axis (*b*) of the cylinder exceeds that acting parallel to the axis (Fig. 4B). The cylinder elongates parallel to *b*, which is the principal symmetry axis of strain. In a rotating quartz pebble the compressive stress acting along any transverse diameter varies from instant to instant; but over a period of time its average value exceeds the compressive stress acting parallel to *b* of the regional movement plan. Thus it could elongate continuously parallel to *b*. This approaches the condition of squeezing normal to *b* which Sander has called *Einengung* (Cf. Turner, 1953, p. 96–99). It is important to rember that in a heterogeneous mass such as a conglomerate-schist complex total strain is the summation of (1) the strain of discrete elemental bodies (*e.g.*, quartz pebbles and micaceous laminae) and (2) relative mutual displacement of such bodies. In the ultimate product of deformation the first of these components may be conspicuous (as in elongated pebbles), while the second component is recorded in more subtle qualities of fabric, notably those pertaining to symmetry.

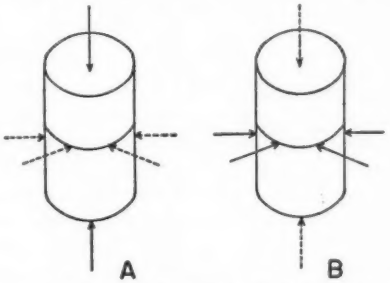

FIGURE 4.—STRESS PATTERNS IN EXPERIMENTAL DEFORMATION

Maximum stress is shown by heavy-lined arrows; minimum stress by broken-lined arrows. A. Compression. B. Elongation.

(4) Correlation of lineation and the trend of aligned pebbles with the *a* axis of movement raises a host of difficulties regarding details of the movement plan. Thus the movement plans developed by Oftedahl and by Brace to explain deformation of conglomerates seem internally inconsistent or improbably complex. Both Kvale (1945, p. 205–207) and Brace (1955, p. 132, 133) appeal to imaginary orienting mechanisms unsupported by experimental evidence. All three explain deformation of quartzite pebbles in terms of mechanical movements alone, and neglect possible recrystallization. Balk was obliged to neglect obvious triclinicity of fabric in attempting to correlate quartz orientation patterns with movements parallel to lineation. These difficulties do not arise if monoclinic fabrics are interpreted as the product of movements normal to the lineation and if the possible influence of repeated deformation or of initial anisotropy of fabric is borne in mind.

SYMMETRY OF EXPERIMENTALLY DEFORMED FABRICS

General Statement

Since Sander and Schmidt first used symmetry as the essential criterion of movement in rock deformation, there have been numerous detailed studies of preferred orientation of crystals in experimentally deformed materials. These provide data wherewith to test the validity of the symmetry concept as applied to highly deformed polycrystalline aggregates. The following are illustrative notes which collectively substantiate the Sander-Schmidt hypothesis as applicable within the field of a small laboratory specimen. The illustrations here selected relate to metals, clays, and marble.

Metallic Aggregates

Patterns of preferred orientation of crystals in deformed metallic aggregates are determined by X-rays and recorded in stereographic projection as pole figures (Cf. Barrett, 1952, p. 170–177). These are particularly valuable for our present purpose in that they clearly illustrate the symmetry of fabrics developed during extreme deformation.

Rolled textures in metal sheets show perfect orthorhombic symmetry (Cf. Barrett, 1952, p. 459–480, Figs. 19–40). The movement plan likewise is orthorhombic. Unless the crystal glide system is known, there is nothing in the fabric pattern to identify any of the three sym-

metry axes as *a*, *b*, or *c* of the movement plan. Balk (1952, p. 430–433) noted that rolled metal fabrics commonly display lineation parallel to the direction of rolling (= *a* of the movement plan). However, this is but one of three binary symmetry axes. It is somewhat analogous to the well-known *a* lineation of some mylonites, but should not be compared with the *b* lineation (sole symmetry axis) of much more common monoclinic fabrics. Pole diagrams for rolled metal commonly show simple or complex girdle patterns. In some of these the girdle axis may be the *a* direction of movement (*e.g.*, Barrett, 1952, p. 478, Fig. 38a); in others it may coincide with *b* (*e.g.*, Barrett, 1952, p. 473, Fig. 33a; p. 474, Fig. 34a); in yet others there may be crossed girdles whose axes are symmetrically inclined to *a* and *b* (*e.g.*, Barrett, 1952, p. 467, Fig. 29b). There is nothing in a girdle axis that necessarily identified it as *b* of the movement plan—notwithstanding which, some writers erroneously attribute such an opinion to Sander (Cf. Strand, 1944, p. 25). However in the special case of a monoclinic fabric (not typical of rolled metals) a girdle axis that is also normal to the sole symmetry plane assumes special significance and can be identified by its symmetry as the *b* axis of movement.

Cold-worked metal fabrics are products of plastic flow without appreciable recrystallization. When heated to a few hundred degrees they become changed or even obliterated by annealing recrystallization. In the former the annealed metal fabric retains a symmetry compatible with that of deformation (Cf. Barrett, 1952, p. 490, 491).

Clay Aggregates

Williamson (1954; 1955) has described patterns of preferred orientation of tabular clay crystals and of rutile and tourmaline needles that develop during deformation of ceramic clays. The orienting mechanism is believed to be relative movement of the solid particles induced by movement in the continuous intergranular film of water. No plastic deformation of the mineral grains is involved. Three kinds of experiment are described: (1) extrusion of a cylinder; (2) rolling of a cylinder; (3) radial spread of a disc. The symmetry of the resulting fabrics (preferred orientation of crystals of clay minerals and of prismatic crystals, *s* planes, and tension joints) is identical to that of the movement plan. Lineation (alignment of prismatic crystals) develops parallel to the *a* direction of movement in the first type of experiment and parallel to *b* in the other two.

Marble

General statement.—The only rock whose deformation has been experimentally investigated in detail over a wide range of geologically possible conditions is a calcite marble from Yule Creek, Colorado. At 150–500° C and confining pressure of 3000 to 5000 atmospheres two gliding mechanisms have been proved for calcite crystals: (1) twin gliding on $\{01\bar{1}2\}$, the sense of displacement of upper layers being upward toward the *c* axis; (2) translation gliding on $\{10\bar{1}1\}$ (Turner, Griggs, and Heard, 1954). During deformation of marble under the same conditions, some grains respond to stress by twinning on $\{01\bar{1}2\}$, others by translation on $\{10\bar{1}1.\}$ Every grain tends to rotate bodily in the sense opposite to that of internal gliding. The nature and results of these movements have been discussed in detail (Turner, Griggs, Clark, and Dixon, 1955). Especially significant for the present purpose is symmetry of stress, strain, and fabric (preferred orientation of calcite) in different experiments.

Symmetry of stress.—When, as normally happens, opposite ends of the marble cylinder remain coaxial during deformation, symmetry of stress is axial. External pressures acting in all directions normal to the cylinder axis are equal. In compression experiments they are less, and in extension experiments greater, than the pressure acting parallel to the cylinder axis (Cf. Fig. 4).

Symmetry of strain.—The shape of the cross section of the deformed cylinder expresses symmetry of strain, which in turn reflects the summation of internal movements in the process of deformation. It is determined by the relative orientation of the system of applied force and the initial fabric of the marble. In the experimentally investigated Yule marble the initial fabric is characterized by a concentration of *c* axes at high angles to a single foliation plane. Its symmetry is essentially axial, the principal symmetry axis being normal to the foliation.

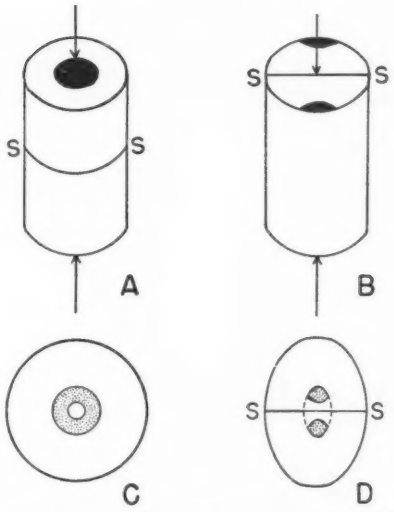

FIGURE 5.—SYMMETRY OF COMPRESSIONAL STRAIN
Full black = concentration of *c* axes of calcite in initial fabric, as projected on end of cylinder. SS = initial foliation. A. Compression normal to foliation. B. Compression parallel to foliation. C. Cross section of strained cylinder after compression normal to foliation. Stippled area = concentration of *c* axes of calcite. D. Cross section of strained cylinder after compression parallel to foliation. Stippled areas = concentration of *c* axes of calcite.

section is an ellipse. The length of one principal diameter, in the plane of initial foliation, has changed but slightly (Cf. Table 1). The other diameter, coinciding with the original direction of preferred orientation of *c* axes, has notably lengthened in compression experiments or shortened in extension experiments. This type of strain implies that the internal movements concerned were mainly directed parallel to the plane containing the cylinder axis and mean trend of *c* axes in the calcite grains.

In a few of our experiments the symmetry of strain is monoclinic. This happens either (1) where the loading in a compression experiment is slightly eccentric so that the stress system has monoclinic symmetry (Griggs, Turner, Borg, and Sosoka, 1953, p. 1340, Pl. 5); or (2) where the axis of compression or elongation is inclined at 45° to the initial foliation (*d* cylinders, Griggs, *et al.*, 1953, p. 1335, Pl. 2, C and D, Fig. 7, C and F). In both strain tends to be concentrated within a diagonal shear zone somewhat comparable with a kink band. This zone is rotated with respect to the less deformed ends so that symmetry of strain for the specimen as a whole is monoclinic. Within the limited zone of shear only sense of shear marks the

TABLE 1.—VALUES OF STRAIN IN DEFORMED MARBLE CYLINDERS*
Initial diameter = 0.5 inches

Experiment	Temperature °C	Orientation relative to foliation F	Maximum strain per cent shortening or elongation	Diameters of cross section (inches)		Symmetry
409	400	Compr. ⊥ F	39.9	.660	.677	Axial
444	400	Elong. ⊥ F	30.1	.417	.422	
477	500	Elong. ⊥ F	93	.363	.368	
443	400	Compr. ∥ F	40.7	.581	.745	Ortho-rhombic
459	500	Elong. ∥ F	118	.415	.285	

* From data supplied by D. T. Griggs

When the axis of compression or of extension is normal to the initial foliation, the unique symmetry axis of stress coincides with that of the original fabric. The deformed cylinder remains approximately circular in cross section, *i.e.*, the symmetry of strain likewise is axial (Fig. 5, A and C). However, when the axis of compression or extension is parallel to the initial foliation, the symmetry of strain is orthorhombic (Fig. 5, B and D). The cross movement plan as monoclinic rather than orthorhombic. For such a small field the strain is essentially orthorhombic.

Symmetry of fabric.—In highly deformed cylinders the symmetry of the newly deformed fabric corresponds precisely with that of strain. The *c* axes of calcite grains become sharply concentrated at angles of 10–30° to the axis of compression or at 70–80° to the axis of elongation (Fig. 6, B and C). Where symmetry of strain

is polar the calcite axes tend to be evenly spread in an annular zone around the cylinder axis, so that the *c*-axis orientation diagram also has polar symmetry (Fig. 7A). The orientation diagrams for prominently twinned $\{01\bar{1}2\}$ lamellae conform to this general symmetry: there is a single lamella maximum coinciding with the pole of the compression axis, or a lamella girdle around the extension axis. Sharply contrasted as regards symmetry are orthorhombic orientation diagrams for cylinders with orthorhombic strain. In these the *c* axes are concentrated in a pair of maxima at about 15° to the compression axis (Fig. 7 B) or in a pair at 70° to the extension axis (Fig. 7 C).

The behavior of individual grains, variously oriented in the stress field, has been deduced from preferred orientation patterns in the deformed marble fabrics and from internal structures in the calcite grains themselves (Turner, Griggs, Clark, and Dixon, 1955). Each grain is progressively reoriented toward an ideal stable orientation with its *c* axis at 20–30° to the compression axis or at 70–80° to the extension axis. The path followed by the *c* axis of any grain, and hence its final position in the cone around

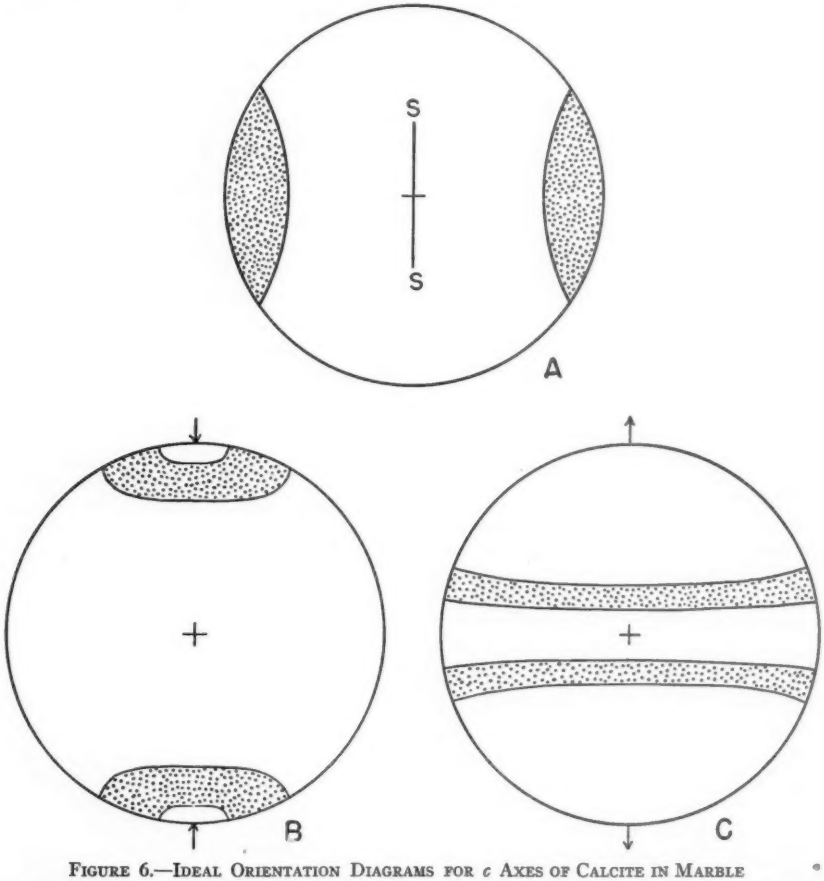

FIGURE 6.—IDEAL ORIENTATION DIAGRAMS FOR *c* AXES OF CALCITE IN MARBLE

A. Initial fabric of Yule marble. Stippled area = zone of concentration of *c* axes. SS = foliation. B. Compression fabric. Stippled area = stable orientation of *c* axes. C. Extension fabric. Stippled, area = stable orientation of *c* axes.

the principal stress axis, is determined by the initial orientation of the crystal in relation to the axis of applied stress. Where the axis of compression or extension coincides with the mean orientation of *c* axes (Figs. 6, B and C, 7 A) these *c* axes may reorient themselves by radial migration in almost any direction. Symmetry of motion and hence of ultimate orientation of *c* axes is axial with reference to the stress axis. Where the axis of compression or of extension is normal to the mean trend of *c* axes (Fig. 7, B and C), the available paths of migration are limited to the vicinity of the arc that connects these two directions. Symmetry of motion and of the resultant orientation pattern of *c* axes is orthorhombic.

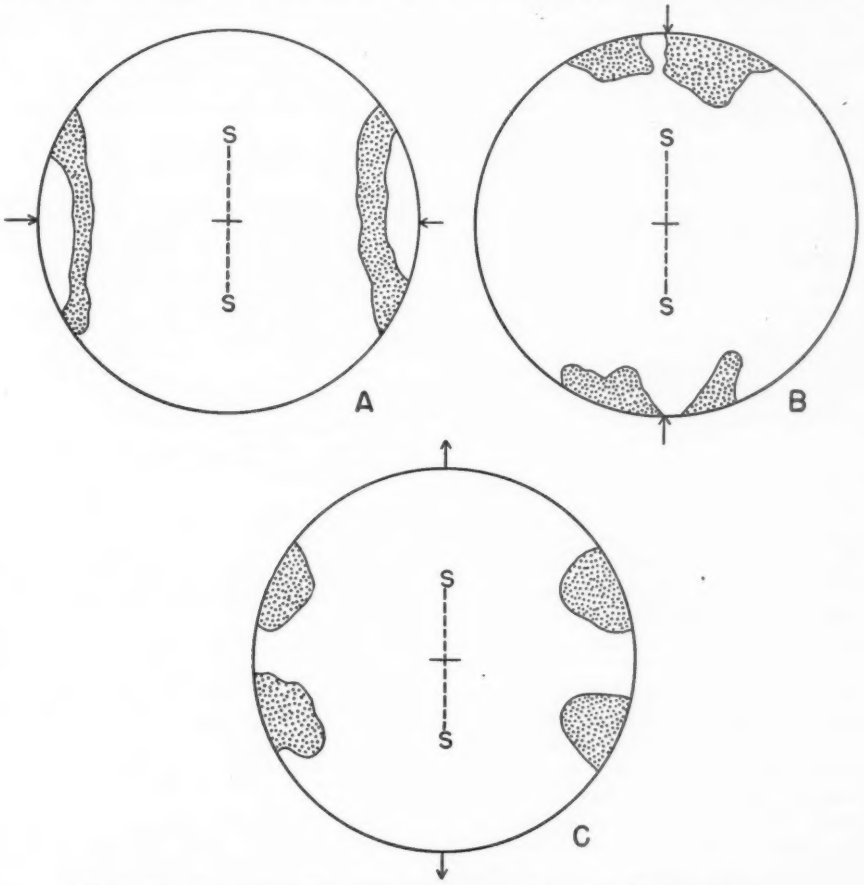

FIGURE 7.—ORIENTATION DIAGRAMS FOR *c* AXES OF CALCITE IN EXPERIMENTALLY DEFORMED MARBLE

SS = initial foliation. Arrows show applied force. Stippled areas are bounded by the 5-per-cent contour drawn for a plot of 100 *c* axes. A. Specimen 409, shortened 39.9 per cent, 400° C, 3000 atmospheres. B. Specimen 443, shortened 40.7 per cent, 400° C, 3000 atmospheres. C. Specimen 459, elongated 118 per cent, 500° C, 5000 atmospheres.

The details of this movement plan, although supported by many observations and in harmony with current concepts of deformation in metallic aggregates, are hypothetical. Whether they are substantiated or rejected in the light of future work, three significant

facts related to the subject of this paper remain established:

(1) Symmetry of fabric in deformed marble is identical with symmetry of strain.

(2) Symmetry of strain reflects symmetry of the over-all movement plan.

(3) Strain, and hence symmetry of the strained fabric, is determined by the stress system in relation to an initially anisotropic fabric. Considering that the degree of anisotropy (indicated by preferred orientation of *c* axes) in the original fabric of Yule marble is not high, it is remarkable how persistent is its influence on strain and on the resultant fabric. The influence of much stronger anisotropy typical of bedded sediments and conspicuously foliated rocks must be even more powerful in metamorphic deformation under natural conditions.

Conclusion

The aim of this paper is to present a reasoned conclusion, open to future revision, as an alternative to that expressed in recent years by Strand, Kvale, Anderson, and others. They hold that regional lineation normal to the symmetry plane of a consistently monoclinic fabric commonly represents the principal direction *a* of movement in deformation. The alternative opinion here presented is that the most satisfactory basis for kinematic interpretation of the fabric of deformed rocks is the symmetry concept originally put forward by Sander and Schmidt. This concept is supported by an impressive array of experimental evidence. It is also completely in harmony with results obtained from many detailed field and laboratory studies of fabrics of naturally deformed rocks (*e.g.*, Flinn, 1952; Elwell, 1955). When kinematic deductions based on symmetry conflict with conclusions reached on other grounds, we are as yet not justified in rejecting the evidence afforded by symmetry. The apparent conflict may at times be based upon misapprehension. For example, contrary to rather prevalent opinion, existence of regional or local lineation parallel to the *a* movement direction, or even the presence of *a* girdles in orientation diagrams for crystal directions (*e.g.*, *c* axes in quartz), is in no way incompatible with the symmetry principle. On the other hand, it is significant that where there is only one plane and axis of symmetry in a monoclinic rock fabric these are also the symmetry plane *ac* and axis *b* of the movement plan. Where conflict is real, the evidence from fabric symmetry may be used to explore the expectable complexity of movements participating in prolonged regional deformation of rocks.

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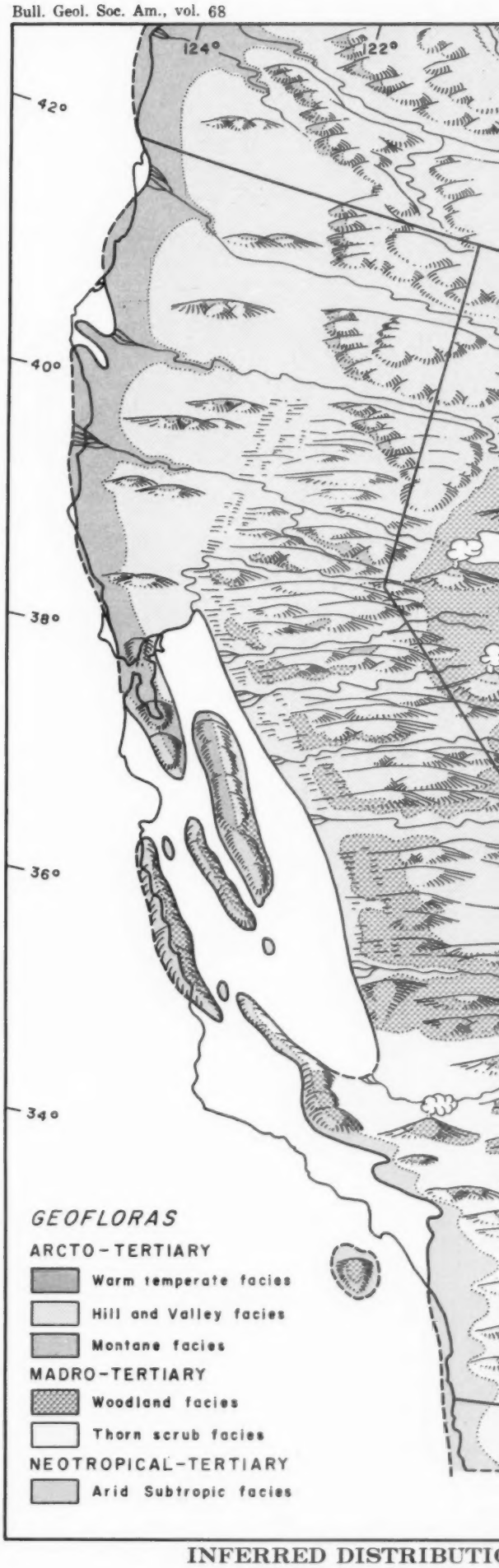

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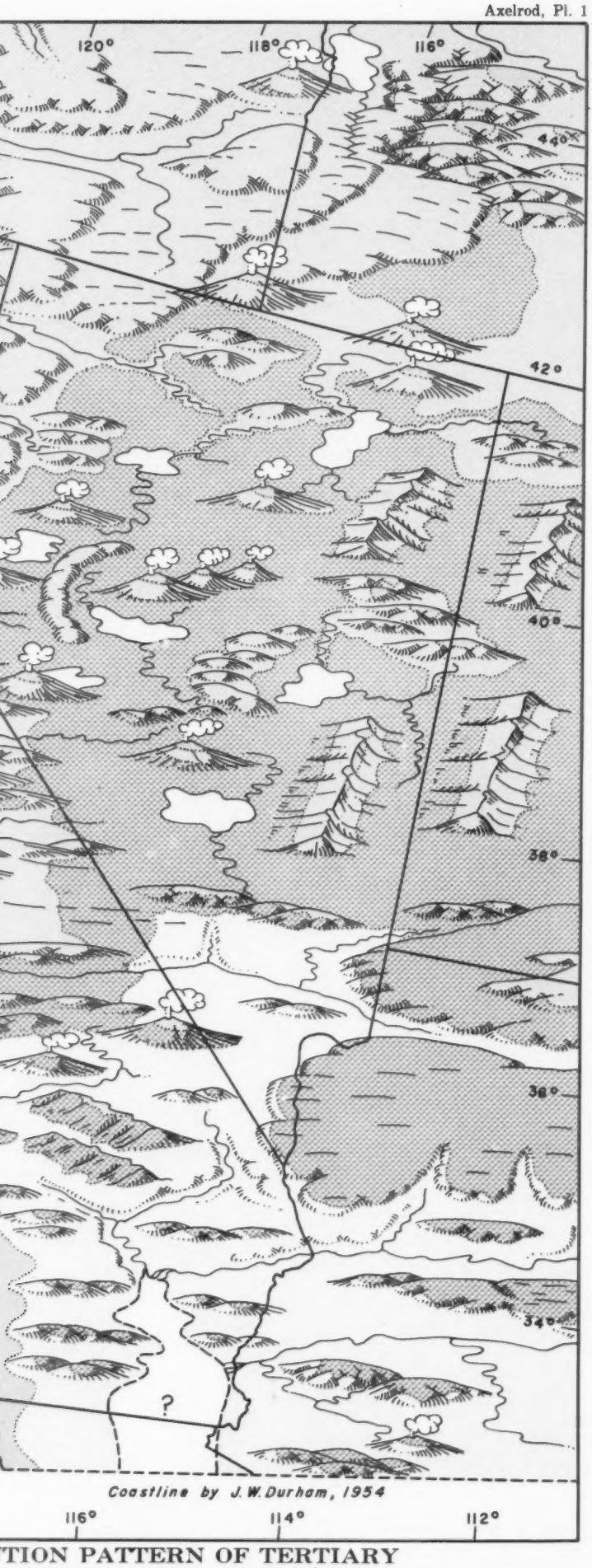

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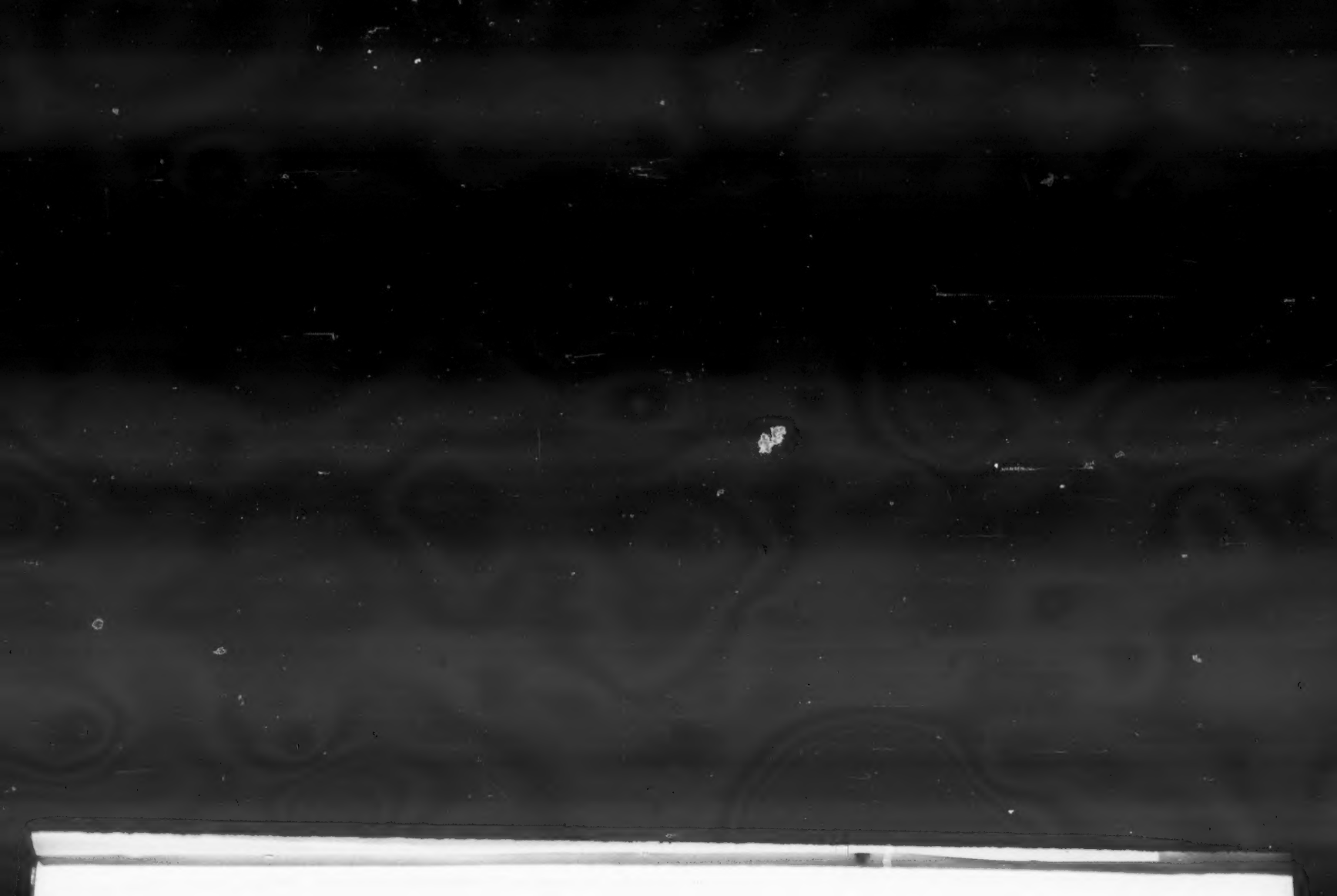

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LATE TERTIARY FLORAS AND THE SIERRA NEVADAN UPLIFT

BY DANIEL I. AXELROD

ABSTRACT

Twenty-one Mio-Pliocene floras distributed from sea level in west-central California, across the Sierra Nevada and into western Nevada, are preserved in volcanic sediments occurring either (1) at or near the base of the andesitic Mehrten formation and its correlatives east and west of the Sierra Nevada, or (2) in the underlying rhyolitic Valley Springs formation and its close equivalents in western Nevada and coastal California. Since the floras are related to living communities whose topographic and climatic relations can be measured, they provide a basis for estimating the altitude of the Sierra Nevada and of west-central Nevada at the inception of vulcanism during Mio-Pliocene time. Local, provincial, and regional comparisons indicate that there were then only moderate east-west differences in vegetation, climate, and topography across this area where environmental diversity is marked today.

A number of paleogeographic inferences are made: *Geologic*—Just before late Tertiary vulcanism the Sierra Nevada was a broad ridge with its summit near 3000 feet. It was incised by 2- to 4-mile wide, 1000-foot-deep valleys, some of which drained western Nevada. The lowlands of western Nevada had an average elevation of 2000–2500 feet. Two major components of late Cenozoic diastrophism are recognized in the Sierra Nevada. Faulting accounted for an average maximum displacement of 3500 feet in this section of the range, and regional warping ranged from 1800 feet at Donner Pass to 3000 feet at Carson Pass 50 miles south. *Climatic*—Building up the Sierran ridge from 6000 to 8000 feet by vulcanism and diastrophism in the later Cenozoic (1) reduced temperature along the summit so that arctic-alpine and subalpine forest environments came into existence, and (2) lowered rainfall sufficiently over the lowlands to leeward to account for the present steppe (sage) and desert regions. *Biologic*—Plants (and animals) from contiguous milder environments gradually were invading these new zones during the later Cenozoic and have continued to evolve there.

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The Problem

The northern and central sections of the Sierra Nevada and the adjoining parts of Nevada are blanketed chiefly by upper Tertiary volcanics and associated sediments. An important hiatus separates them from the underlying section, most of which is a basement of metamorphic and granitic rocks, though Eocene and Oligocene rocks are represented locally. The upper Tertiary volcanics locally exceed 3500 feet in thickness and rest on a surface of low to moderate relief. The section has been folded and faulted, with total displacement along the north-central portion of the Sierran scarp averaging 3500 feet during late Pliocene and Pleistocene times. Therefore, what was the elevation of the range and of the easterly bordering area just before the extensive rhyolitic and andesitic eruptions near the close of Miocene time, following a long period of erosion?

This problem has earlier attracted the interest of geologists. Lindgren (1911), in particular, discussed the evidence provided by the gradients of the Tertiary Sierran stream channels and concluded that the summit of the range was between 5000 and 7000 feet in Miocene time. This excessive estimate resulted from three misconceptions. First, Lindgren thought that the broad channels containing the auriferous gravels (by which he calculated the stream gradients) and the overlying volcanics were all the same age, namely Miocene. It is clear now that most of the older gravels are part of the Ione formation of early Eocene (Capay) age (Allen, 1929; MacGinitie, 1941), and that a major hiatus separates them from the volcanic section comprising the Valley Springs and Mehrten formations which range from Mio-Pliocene well into Pliocene time. Lindgren's data on the gradients of the stream channels thus relate to the profiles of the Eocene rivers, not to those of late Tertiary age. As shown by Hudson (1955, p. 855), a discordance between the Eocene and Mio-Pliocene rocks indicates that some deformation had taken place during this interval, at least locally in the range.

Secondly, Lindgren's data on the gradients of the channels are subject to error. His sections suggest that both folding and faulting took place within the range in subsequent times, yet in determining the original stream gradients he did not compensate for structure. Following the views of King, LeConte, and others, he believed that the range had reacted substantially as a rigid block (Lindgren, 1911, p. 48). Hudson (1948; 1951; 1955) has shown, however, that folding and faulting were important in the late Tertiary history of the range, not only in the summit section, but also within the range.

Finally, by a peculiar line of reasoning which was colored in part by his erroneous comprehension of the structure of the range, Lindgren concluded that the streams started on a drainage divide that lay a few miles west of the present one. Actually, some of the gold-bearing channels which he thought headed in the range extended eastward into Nevada (Reid, 1911; Gianella, 1936). His estimate of a 5000- to 7000-foot crest for the central to northern Sierra Nevada in the early Tertiary is too high, as shown by the regional relations of the Eocene floras in the far West.

Matthes (1930) also made an estimate of the height of the central part of the range in connection with his study of the Yosemite area. He pointed out that an old erosion surface (the broad valley stage) carved into the crystalline rocks could be traced from Yosemite Valley into the summit region, as well as northwestward in the range for 40 miles where it lay covered by the andesitic Mehrten formation near Tuolumne Table Mountain. He dated this surface late Miocene (here considered Mio-Pliocene), for at Table Mountain plants (Table Mountain flora) and mammals (Springfield Shaft local fauna) of this age occur at the base of the volcanic section. Matthes then pointed out that the erosion surface had an elevation of approximately 4000 feet in the Yosemite summit region in Mio-Pliocene time.

This figure was obtained by asserting that uplift at the end of the Miocene added 3000 feet to the range and that faulting at the close of the Tertiary increased the height of the block by an additional 6000 feet. The sum of these subtracted from the present summit level of 13,000 feet gives the figure of 4000 feet.

A number of objections concerning Matthes' interpretations have been raised. Some geologists have expressed skepticism that the upland surface in the Yosemite area corresponds to that on which the Mehrten formation was deposited. They point out that there is no evidence that the Yosemite area was covered by andesite, and hence the erosion surface may have been developing since Miocene time, or earlier; in any event, the present surface is of later Cenozoic, not of Mio-Pliocene age, for it is not covered by volcanics. Others have examined the region and doubt that there is an old erosion surface, at least in Matthes' sense.

Of greater importance to the problem of elevation, however, is that Matthes cited no facts in support of Mio-Pliocene faulting in the Yosemite area. Actually, no such evidence exists there, for the rocks in the uplands are pre-Tertiary crystallines and provide no data for estimating the amount of faulting at any time in the Tertiary. His estimate of 3000 feet uplift of the close of the Miocene apparently was derived from the work on the Truckee quadrangle which led Lindgren (1897) to believe that there was major pre-andesite faulting in the region. Although there is evidence of faulting during the short interval between the accumulation of the Valley Springs and Mehrten formations, as at Donner Pass (Hudson, 1948; 1951), it was relatively minor. This Mio-Pliocene faulting does not account for the Tahoe graben, as Lindgren supposed; it is much younger. Recent work by Garniss Curtis (September 1955, personal communication) in the area from Carson Pass to Ebbets Pass, 50 to 65 miles south, has failed to demonstrate the existence of major pre-andesite faulting in that broad region. To suppose that large-scale faulting took place in the Yosemite region another 70 miles south in Mio-Pliocene time seems unwarranted. Thus, whereas Matthes' estimate of the altitude of the summit region is not greatly different from that provided by the paleoecologic analysis presented below, the method by which his conclusion was reached is unsound.

Hudson (1955) measured post-Eocene deformation in the Sierra Nevada by using the gradients of the Tertiary Yuba River which are marked by the old auriferous gravels that can be traced from sea level into the summit area. Rather than relying on the gradient of the complete stream course, which has been warped and faulted, Hudson chose several triplets (three consecutive reaches of the channel) for study. The method, he points out (1955, p. 850),

> "*. . . does not yield the actual original gradient of a triplet of reaches, but what the original gradient would have been if the three reaches had a common gradient and lay on a block that remained rigid during tilting.* In order to learn the probable effect of lack of equality of the original gradients of the members of the solved triplets, one must first decide upon the probable amounts of variance from equality. This was done by studying parts of the modern Yuba river where the gradients approximate those for the eight usable triplets of the ancient stream."

Hudson concluded that the Sierran summit was 5500 feet in the early Eocene (1955, Fig. 2B) and that total uplift of the summit region at Donner Pass has been only 1986 feet since that time. These results do not agree with the evidence of altitude supplied by the late Tertiary floras in the near-by Sierra Nevada and adjacent Nevada. They suggest that the range was the site of a broad low ridge at the close of Miocene time, and that in the Donner area uplift has been on the order of 5300 feet since the middle Pliocene.[1]

Acknowledgments

This study is one of the results of a long-term investigation of late Tertiary paleobotany in the far West. The project has been supported by the Carnegie Institution of Washington, the National Research Council, the Committee on Research, University of California, Los Angeles,

[1] Hudson and the writer have discussed their different interpretations more than once. In his latest letter (Oct. 12, 1955) he states: "One or the other of us is wrong, or both of us are wrong in greater or less degree. I think your course should be, not to try to reconcile the results, but to state that the paleobotanic evidence indicates that my estimate is too small."

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In considering some of the topographic relations of climate which are presented, the writer has been helped greatly by discussions with Harry P. Bailey and Harry D. MacGinitie, both of whom have a keen sense of modern climates and their relation to those of the past.

The manuscript has been read critically by George A. Bartholomew, Ralph W. Chaney, Garniss Curtis, Erling Dorf, Cordell Durrell, Carl Epling, Frank S. Hudson, Harry D. MacGinitie, William C. Putnam, and Edward L. Winterer. They have added to its clarity and made several important suggestions which have been incorporated in the discussion.

Method

Fossil plants have been used repeatedly to interpret the nature of ancient topography, especially during Cenozoic time because most Cenozoic plants closely resemble living species whose present requirements in terms of terrain and climate can be analyzed. Although one plant may provide evidence of past environment within broad limits, sounder results are obtained by comparing fossil floras with modern communities which resemble them most closely. In this way details of relief and climate surrounding a basin of deposition can be outlined with reasonable accuracy. Comparisons of floras of similar age over wider areas permit inferences of broader paleogeographic significance.

The interpretation of ancient relief, particularly its bearing on altitude, falls into two categories—local and regional. In a local problem the difference in elevation between the environment indicated by the fossil flora and the modern flora at the site of deposition is determined as nearly as possible. For example, Puri (1947) studied the rich Pleistocene Liddarmarg flora which occurs at 10,600 feet in the foothills of the Himalayas. It contains numerous warm-temperate or subtropical plants whose upper altitudinal limits now lie from 4000 to 6000 feet lower than the basin of deposition. He concludes that since the beds were deposited the range has been uplifted on the order of a mile.[2] Berry (1917; 1922a; 1922b; 1938, p. 49–51) described several fossil floras from the mountains of Bolivia which occur at elevations of more than 12,000 feet. They contain subtropical and tropical plants similar to those living today at much lower altitudes. He concluded that the Andes must have been elevated a mile or more in this area during the late Cenozoic, bringing these localities into a cold high-altitude desert climate where the vegetation is very different. MacGinitie (1953) showed that the rich Florissant flora, preserved in lake beds in central Colorado at an elevation of 8500 feet, is a mixtu of plant communities whose general ecologica relations suggest that this Oligocene basin was no higher than 3000 feet and was probably lower.

In regional studies, comparisons of fossil floras of similar age that now lie on opposite sides of a high mountain range may demonstrate that the present intervening barrier could not have existed when the floras were living. Berry (1922c) pointed to the marked differences between the present temperate rain-forest climate on the coastward slopes of the high Andes in southern Chile and the cold desert climate of Patagonia to leeward. He found that the warm-temperate flora associated with the Eocene Concepcion-Arauco coal beds in southern Chile (Berry, 1922c) resembled the Rio Pichileufu flora across the Andes in Patagonia (Berry, 1938) and concluded that the range was not yet a major climatic barrier when these floras were living (Berry, 1938, p. 53). In the same way, Chaney (1938a; 1938b) noted that the Cascades now separate widely different environments in Oregon: humid-temperate conifer forests densely blanket the windward side of the range and open semiarid sage and desert communities occur over the lowlands to leeward. By contrast, the Eocene subtropical lowland forests of interior Oregon are closely related to those of the western coastland. Such similarities of vegetation

[2] In this example, Puri did not consider the possibility that during the Pleistocene climatic change alone may have accounted for the shift of a lowland flora into the uplands, or vice versa.

and the climate inferred suggest that the Cascade range did not exist as a high topographic barrier in Eocene time.

Comparisons of floras of similar age on both sides of a mountain range may show sufficient differences in vegetation, and the climates they suggest, to indicate that the range was already beginning to rise. Chaney (1944b, 358–360) compared the early Pliocene Troutdale and Dalles floras from the windward and leeward sides of the Cascades in northern Oregon at the latitute of the Columbia River. The differences between them suggest precipitation was approximately 10 to 15 inches less in the lowlands to the east of the ancestral Cascades. This is paralleled by modern relations in the Coast Ranges of California, where ridges 1500 to 1700 feet high account for similar differences in related vegetation and climate on their windward and leeward slopes. Thus Chaney suggests that the central Cascades may have had a similar elevation in the early Pliocene.

These local and regional methods of analysis are used here to estimate the altitude of the Sierra Nevada and the plateau which lay to the east of it during the late Tertiary. First for analysis is a group of Mio-Pliocene floras (early Clarendonian, Neroly-Cierbo, late Mohnian-Delmontian ages) that extend from sea level in coastal California across the Sierra Nevada and into western Nevada, supplemented by data from floras of similar age to the north and south. They will indicate the relative altitude of the region at the outbreak of vulcanism in Mio-Pliocene time. A number of Pliocene floras in the area are then referred to in the section on paleogeography, for they have an important bearing on later topographic, climatic, and biologic changes in the region. Before discussing the paleobotanic evidence, however, the stratigraphic occurrence of the floras must be summarized briefly.

Stratigraphy

In recent years studies of numerous late Tertiary floras in west-central California and in the Sierra Nevada have been completed (Chaney, Condit, and Axelrod, 1944); Axelrod, 1950a); studies of other floras are in their final stages (Axelrod, in press; MSS), and others have been collected and studied sufficiently so that preliminary conclusions can be drawn with respect to their ages and the environment in which they lived. Figure 1 shows that more than 35 floras of middle and late Tertiary age occur in the area from sea level in central California across the central Sierra Nevada into western Nevada, and that others lie to the north and south. More than 20 of these floras lived in the late Miocene or early Pliocene. There is still much overlap in the usage of these terms on the Pacific Coast, though the time dealt with here can be delimited precisely in provincial ages. Many of the floras occur in rocks that have yielded mammalian faunas of early Clarendonian age, which most vertebrate paleontologists regard as early Pliocene (Wood *et al.*, 1941). Others come from beds associated with marine megafossils which represent the Cierbo and Neroly "stages" of the marine molluscan sequence and are commonly considered late Miocene (Weaver *et al.*, 1944), though they have also yielded early Pliocene mammals. Some of the floras are in rocks containing marine microfossils that range from the Mohnian into the Delmontian, regarded generally as late Miocene and Mio-Pliocene, respectively, (Weaver *et al.*, 1944). These rocks have also yielded late Miocene mollusks and in some places interfinger with beds containing early Pliocene mammals. This diversity in the usage of epoch terms on the Pacific Coast is due in part to correlation problems and in part to the fact that the type sections of the Miocene and Pliocene are not continuous. Described from different areas, they are separated by a hiatus that is apparently equivalent to the European Pontian and Sarmatian stages, and to the Barstovian (upper), Clarendonian, and Hemphillian North American mammalian stages. Since there is no agreement on where the Mio-Pliocene boundary should be placed, the floras studied are considered Mio-Pliocene. Their temporal relations are summarized in Figure 2.

There is still much confusion in recent literature concerning the age and stratigraphic relations of the Tertiary rocks in the Sierra Nevada. Since the present problem involves the regional relations of some of these forma-

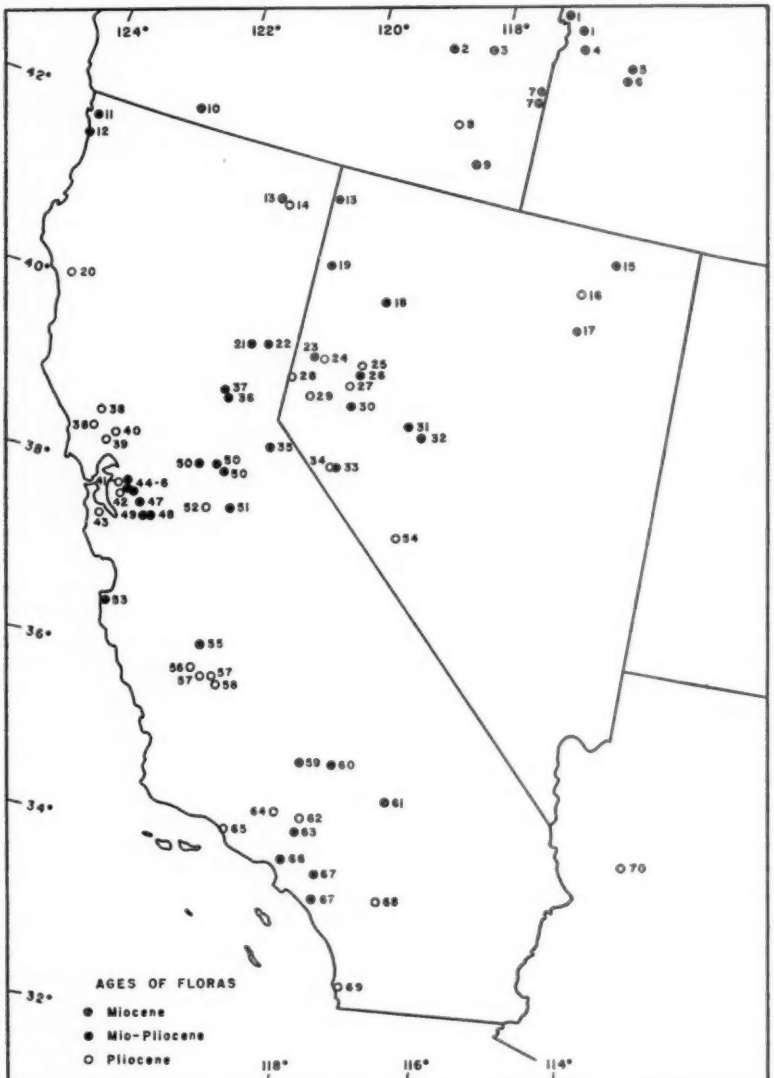
124°
122°
120°
118°
42°
40°
38°
36°
34°
32°
118°
116°
114°
AGES OF FLORAS
Miocene
Mio-Pliocene
Pliocene

FIGURE 1

tions, the Tertiary stratigraphy in the central and northern parts of the range is summarized here (Fig. 3). The diverging lines for the epoch boundaries represent current opinions of the relation of the west-coast stages to the European section. The post-Paleocene epoch boundaries drawn by the mammalian paleontologists fall at the bases of the Wasatchian, Chadronian, Arikareean, and Clarendonian stages, and at the top of the Blancan. The molluscan correlations with the European epochal boundaries are shown by the lines on the right side of the epoch column: the boundaries are at the bottom of the Capay, in the upper third of the Keasey, the middle of the Blakeley, the base of the Jacalitos, and the top of the San Joaquin.[3]

It is apparent that the Mio-Pliocene floras of the Sierra Nevada were living just before or during the early eruptions of the late Tertiary andesites which are termed the Mehrten formation in the Sierra Nevada and the Kate Peak to the east. This is an important stratigraphic position, for in the Sierra Nevada the base of this andesite section at a number of localities is accurately dated as early Clarendonian by mammalian faunas (*Springfield Shaft*— Marriam and Stock, 1933; *Two-Mile Bar*— Stirton and Goeriz, 1942) and floras (*Table Mountain*— Condit, 1944b; *Remington Hill*— Condit, 1944a; *Forest*— Condit, 1944a, p. 38; *Mohawk*— Axelrod, 1944, p. 220–221). The late Tertiary andesites, which were erupted chiefly in the upper parts of the range and to the east in Nevada, were transported to basins which lie farther east and west. Significantly, a fauna of early Clarendonian age occurs at Coal Valley in western Nevada (Stirton, 1939b, p. 634), where it is associated with the lowest Kate Peak andesites (Axelrod, in press). The andesites transported westward across the Valley of California have been found and recognized in the San Pablo (Neroly) formation at Corral Hollow (Louderback, 1924; Huey, 1948), where they interfinger with a marine section. A large flora occurs in the andesitic sandstones at Corral Hollow (Lesquereux, 1883; Condit, 1938); marine megafossils of Neroly age and marine microfossils of Mohnian-Delmontian age are also represented in this section, as are mammalian fossils of early Clarendonian age (Stirton, 1939a, p. 364–365). Thus the lower parts of the Mehrten and Kate

[3] The 18th International Geological Congress recommended that the base of the Pleistocene be placed at the bottom of the Villafranchian; thus on the West coast the base of the Pleistocene probably would fall near the middle of the Blancan and San Joaquin (Fig. 3).

FIGURE 1.—GEOGRAPHIC OCCURRENCE OF MIDDLE AND LATE TERTIARY FLORAS IN CALIFORNIA, NEVADA, AND BORDERING AREAS

Fossil Floras

1. Lower Idaho
2. Stinking Water
3. Beulah
4. Payette
5. Idaho City
6. Thorn Creek
7. Succor Creek
8. Alvord Creek
9. Trout Creek
10. Grizzly Peak
11. Wimer
12. Crescent City
13. Upper Cedarville
14. Alturas
15. Copper Basin
16. Humboldt
17. Elko
18. Rabbithole Springs
19. Juniper Valley
20. Wildcat
21. Monte Cristo
22. Mohawk
23. Pyramid
24. Quail Creek
25. Truckee
26. Chloropagus
27. Hazen
28. Verdi
29. Chalk Hills
30. Fallon
31. Middlegate
32. Buffalo Canyon
33. Aldrich Station
34. Coal Valley
35. Carson Pass
36. Alleghany
37. Remington Hill
38. Sonoma
39. Petaluma
40. Napa
41. Mulholland
42. Black Hawk Ranch
43. Santa Clara
44. Diablo
45. Loma Ranch
46. Bailey Road
47. Altamont Pass
48. Corral Hollow
49. Cierbo
50. Valley Springs
51. Table Mountain
52. Oakdale
53. Carmel
54. Esmeralda
55. Temblor
56. Jacalitos
57. Etchegoin
58. San Joaquin
59. Tehachapi
60. Ricardo
61. Barstow
62. Anaverde
63. Mint Canyon
64. Piru Gorge
65. Pico
66. Modelo
67. Puente
68. Mount Eden
69. San Diego
70. Wikieup

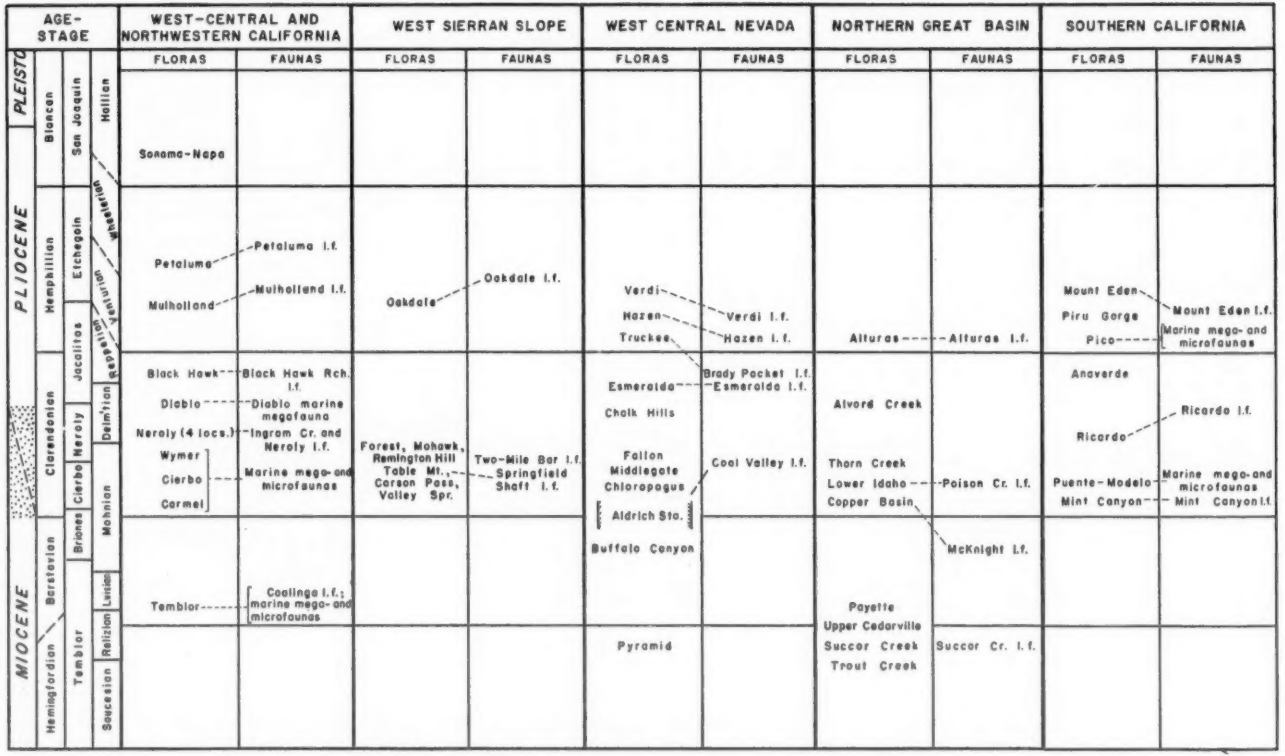

FIGURE 2.—RELATIVE AGES OF MIDDLE AND LATE TERTIARY FLORAS

Occurrence of marine and nonmarine faunas in the section is indicated in the right-hand column for each region; dashed lines show their stratigraphic positions relative to the floras.

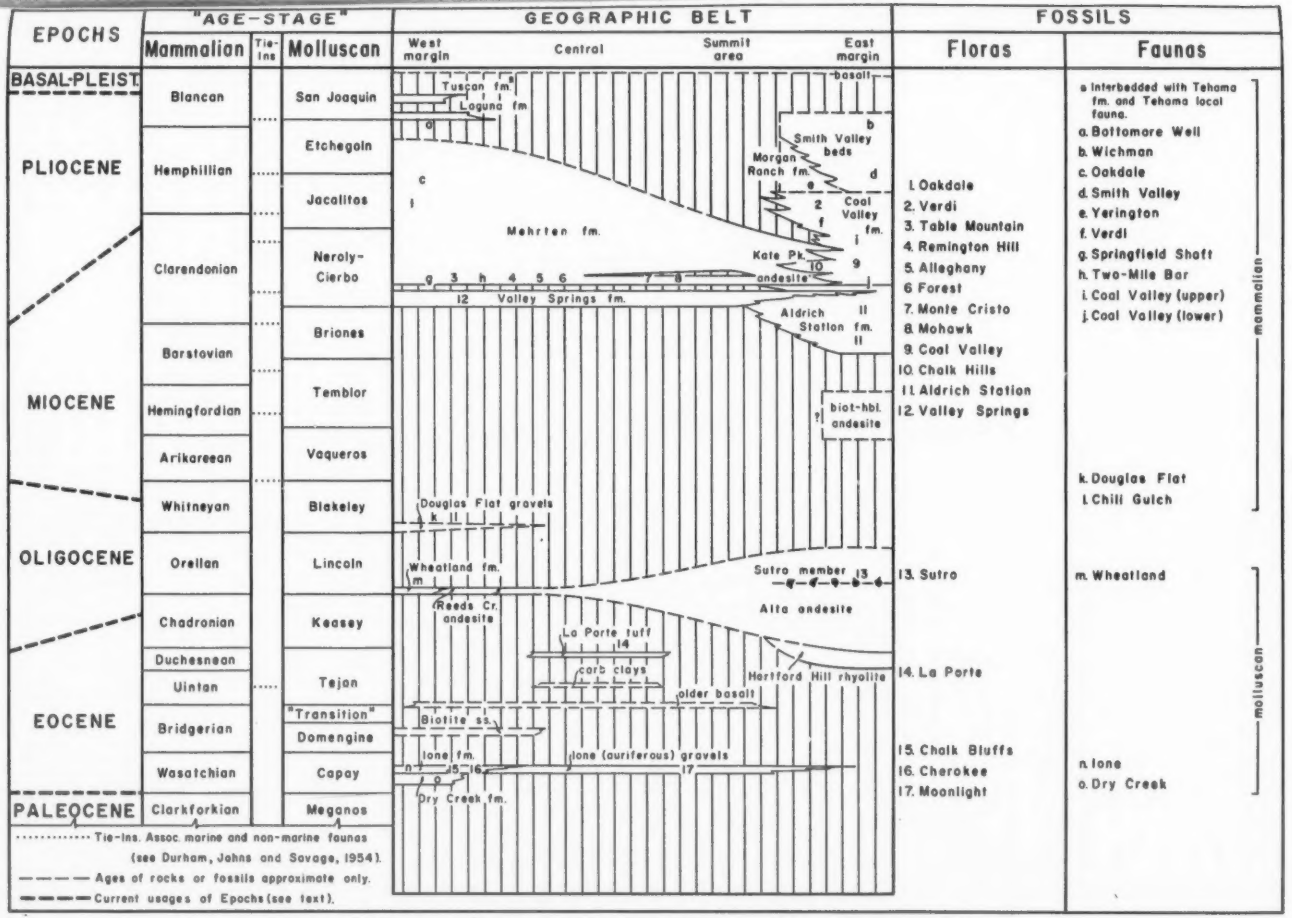

FIGURE 3.—TERTIARY STRATIGRAPHY OF THE CENTRAL AND NORTHERN SIERRA NEVADA

Peak formations in the Sierra Nevada and western Nevada, and the correlative andesitic sediments lying farther east in Nevada and to the west in coastal California represent a good lithic unit, and an important time unit.

In the Sierra Nevada the andesites and andesitic sediments which make up the Mehrten formation are at many points underlain by the Valley Springs formation, which includes as much as 500 feet of rhyolite tuff and associated fluvatile material. Although the contact may be disconformable there is no extensive time hiatus in this section because the Valley Springs flora is similar to that from the base of the overlying Mehrten formation (Axelrod, 1944, p. 219). The Valley Springs has a stratigraphic position no less significant than the base of the overlying Sierran andesites. Reconnaissance work in western Nevada demonstrates that rhyolites equivalent to the Valley Springs conformably underlie the lowest Kate Peak (Sierran) andesites at several localities. This includes the Coal Valley area where mammalian fossils of early Clarendonian age (U.C. Mus. Pal., locs. V3939, V4705, V4706) also occur just below the Kate Peak (Sierran) andesite. At this locality they are in the Aldrich Station formation which is a fluvio-lacustrine deposit of rhyolite tuff, pumice, diatomite, and siliceous shale containing the Aldrich Station flora (Axelrod, *in press*). To the north, at the western margin of the Carson Sink, rhyolite tuff beds apparently representing the Valley Springs formation lie conformably below the Kate Peak andesite and grade eastward into a thick lacustrine section dominated by siliceous shale which has yielded the Fallon flora of early Clarendonian age (Axelrod, *in press*). In the marine section of California the Valley Springs has its equivalent in the thick units of rhyolite tuff that are associated with the Cierbo formation in the Corral Hollow area (Huey, 1948) and southward (Anderson and Pack, 1915, p. 96–100), where it underlies the Neroly formation which is there charged with andesitic debris transported westward from the Sierra Nevada. Associated marine megafossils and microfossils indicate that these rhyolite tuffs are of Cierbo and Mohnian age, and a similar age is assigned to the Cierbo flora of the Corral Hollow area.

Figure 4 illustrates in diagrammatic manner the east-west relations of the base of the Sierran andesite (Mehrten formation, Kate Peak andesite) and the underlying rhyolite (Valley Springs formation). It indicates that the Mio-Pliocene floras were living at a time of rhyolite activity, or immediately following it, during the inception of andesitic eruptions.

To interpret more fully the floras that lie along the transect from western Nevada to the Coast Ranges of west-central California, it is necessary to consider floras of similar age to the north and south because they provide data on regional environments which bear on the problem of altitude. That all these floras are of Mio-Pliocene age rests not only on evidence supplied by the plant fossils; in some areas the age is indicated by associated early Clarendonian mammals, and in other regions by molluscan or foraminiferal assemblages of Cierbo-Neroly and Mohnian-early Delmontian ages, respectively. (*See* Fig. 2.) Particularly critical is the fact that a number of these floras occur in sections that are also characterized by rhyolite tuff. Thus the Hog Creek and associated florules from the lower part of the Idaho (=Poison Creek) formation near Weiser, Idaho, occur in a section rich in rhyolite ash and pumice (Buwalda, 1923; Dorf, 1936). The Thorn Creek flora (Smith, 1941) from the region north of Boise, Idaho, is of similar age and occurs in a section containing pumice and rhyolite ash. The lower part of the Humboldt formation in northeastern Nevada, described by King (1878, p. 434–443), which is equivalent to the middle member of the Humboldt discussed by Sharp (1939), includes rhyolite tuffs associated with a fluvio-lacustrine section. The McKnight mammalian fauna from Bone Valley (King, 1878; Merriam, 1914), of late Barstovian age, apparently occurs low in this section. The ash-rich section of the middle Humboldt formation interfingers northward with a thick rhyolite section that centers near Jarbidge, where it contains the Copper Basin flora of late Barstovian or early Clarendonian age. These rhyolites apparently are equivalent to the Owyhee rhyolite of southwestern Idaho, which lies conformably on the Columbia Lavas and grades upward into the Idaho (= Poison Creek) formation. It would appear that the

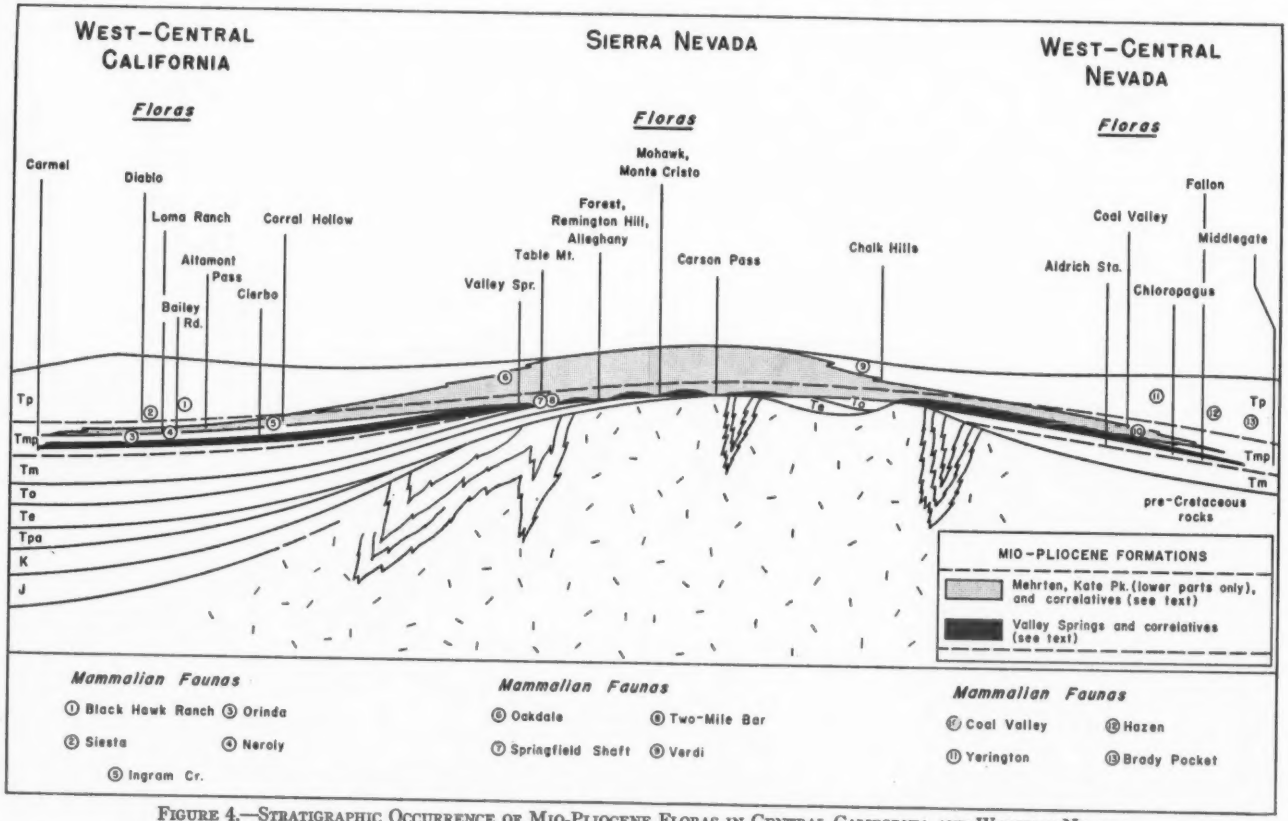

FIGURE 4.—STRATIGRAPHIC OCCURRENCE OF MIO-PLIOCENE FLORAS IN CENTRAL CALIFORNIA AND WESTERN NEVADA

ash-rich part of the Humboldt (*i.e.*, the middle member described by Sharp) is probably at least in part equivalent in age to the Idaho (= Poison Creek), and both may represent the same formation.

The Mint Canyon formation of southern California contains numerous beds of rhyolite tuff in its middle portion where the flora and most of the mammals occur (Jahns, 1940; Axelrod, 1940b). Furthermore, the lower part of the near-by Ricardo has much rhyolite tuff and pumice at the level of the Petrified Forest (Webber, 1933), and the large mammalian fauna of mid-Clarendonian age comes from the overlying section (Merriam, 1919).

It is apparent that rhyolitic volcanics erupted over a wide area in the far West during Mio-Pliocene time and that much of this material was in the form of fine pyroclastics. In this connection it is significant that the diatomites which are so characteristic of the marine section of California also appear to have accumulated at this time when an abundant supply of silica from acid tuffs provided favorable conditions for the optimum development of these siliceous organisms (Bramlette, 1946). As shown by Bramlette, much of the diatomite has been altered to porcelaneous rocks, which, with the diatomite, make up a distinctive part of the Monterey formation. Since rhyolites were erupting contemporaneously in near-by regions eastward, they may have been one of the chief sources for the silica in these marine rocks. Thick sections of rhyolite which are slightly older than those discussed above are known also in the Great Basin and may have provided much of the silica for the earlier Monterey rocks of Luisian-Barstovian ages.

Regional Environments

General Remarks

If the Mio-Pliocene floras of the region are grouped according to similarity, they represent several provinces which differ widely in vegetation and, by inference, in climate. The areas occupied by these Mio-Pliocene environments include the vegetation of three major Tertiary Geofloras.[4] Since these environments have an important bearing on interpreting the altitude of western Nevada and the Sierra Nevada, it is appropriate to review briefly their general nature. Plate 1 has been prepared to facilitate discussion of these Mio-Pliocene environments. It shows the distribution pattern which is inferred to have existed in Mio-Pliocene time, not the actual occurrence of the Tertiary Geofloras. It expresses certain broad conclusions which may be drawn with respect to the areal relations of Mio-Pliocene vegetation.

This map has been constructed on the basis of the relatively few Mio-Pliocene floras known for the broad region shown (*see* Fig. 1), supplemented by inferences derived from somewhat younger and older floras, as well as from geologic and ecologic evidence. A few examples show how these inferences have been used to fill out the map. The western Nevada floras occur in sediments intercalated in volcanic sections typified by flows and pyroclastics ranging from rhyolite to andesite, with basalt also represented. To judge from modern volcanoes, such an occurrence for the floras makes the assumption of moderately high (3000 to 6000 feet) cones near the sites of deposition highly probable. The floras in this region are dominated by live oaks, and forest conifers are subordinate or rare. Comparative studies of modern and fossil plant accumulations have shown that only plants living at the site of deposition are among the dominants, and that those on bordering slopes are rare or absent in both the modern and ancient deposits (Chaney, 1925). In these Nevada floras the inference is that these rare conifers in the lowland floras were living chiefly on the middle and upper slopes of the bordering volcanoes. Similarly, thick boulder beds in the Thorn Creek section of south-central Idaho suggest that mountains bordered the basin of deposition. The fossil plants, preserved in lacustrine shales interbedded with the conglomerates and arkosic sandstones, contain numerous conifers. That they were more abundant on the bordering slopes where climate was cooler is highly

[4] The term Geoflora supplants the earlier term Flora in so far as it applies to the major units of Tertiary vegetation (Chaney and Axelrod, *in press*). For a more detailed discussion of the composition, time-space relations, and evolution of the Tertiary Geofloras, *see:* Chaney (1938a; 1938b; 1944a; 1948), Chaney and Hu (1940), Axelrod (1938; 1939; 1940a; 1950a; 1952).

probable. That a similar distribution prevailed elsewhere in the region wherever mountains were present seems likely in view of the known composition of the floras in the bordering region to northward.

Ecological inferences have been drawn also from the latitudinal and slope relationships of modern vegetation related to the fossil floras. Temperate vegetation grows at higher altitudes southward and at lower altitudes on cooler slopes facing east and north. A similar occurrence is inferred for the Nevada volcanoes and the Idaho hill country during Mio-Pliocene time. By contrast, semiarid vegetation extends northward today on warmer slopes facing south and west and at successively lower altitudes. Oak woodland and chaparral vegetation occupied the lower warmer slopes of central Nevada during Mio-Pliocene time but were replaced at higher levels, where climate was cooler and more moist, by temperate forests (Axelrod, 1940a; in press). Since woodland and chaparral also dominated the interior lowlands of southern California during Mio-Pliocene time, it is inferred that they occupied exposed sites in the intermediate region where no Mio-Pliocene floras are known. As another example, the subtropical Carmel and Puente floras of coastal California are very similar in composition. Although no Mio-Pliocene floras are known in the intermediate area, it is probable that similar vegetation occupied the mild seaward slopes of the coastal archipelago in the intervening region, By using inferences of this sort and depicting them on a schematic map, much can be shown that is purely speculative but is consistent with all the evidence available. Recalling that Plate 1 serves only to express a probable distribution pattern which supplies evidence for interpreting regional climate and topography, we may summarize briefly these Mio-Pliocene environments.

Arcto-Tertiary Geoflora

The Arcto-Tertiary Geoflora was a summergreen temperate forest of deciduous hardwoods and conifers which occupied high northern latitudes during the early Tertiary and spread southward during the rest of the period as climate was cooling. In terms of present affinities, its species were similar to plants now living in three widely separated areas. Members of the West American Element have close counterparts in the conifer forests of western North America. The East American Element included plants similar to the deciduous hardwoods and conifers now found in eastern North America. Nearest relatives of species of the East Asian Element are in the temperate deciduous hardwood and conifer forests of northeastern Asia, a region which is floristically related to eastern North America. The regular association of species of all these elements at numerous localities during Tertiary time demonstrates the holarctic continuity of the Arcto-Tertiary Geoflora, both in time and in space. Yet it was not a homogeneous vegetational type during any part of that period. On the contrary, in each region it regularly displays a diversity of composition in response to local changes in relief and climate, and its successive changes in composition and distribution provide a sound basis for age determination (Chaney, 1936).

During Mio-Pliocene time the Arcto-Tertiary Geoflora was differentiated into three major provinces (Pl. 1). The first was a warm-temperate maritime province which occupied the outer coastal strip from the San Francisco Bay area northward into Washington. It included a number of warm-temperate relicts such as avocado (*Persea*), holly (*Ilex*), magnolia (*Magnolia*), palm (*Sabal*), and others which were surviving near the coast at this later date in a region of mild winter climate. Here they were living with such typical members of the Arcto-Tertiary Geoflora as alder (*Alnus*), birch (*Betula*), persimmon (*Diospyros*), sweetgum (*Liquidambar*), sycamore (*Platanus*), lingnut (*Pterocarya*), swamp cypress (*Taxodium*), and their regular associates. Judging from the requirements of modern species most similar to these fossil plants, rainfall in this coastal area was distributed rather evenly through the year, ranging from 35 to 40 inches at the south and increasing gradually northward along the coast. Summers presumably were warm and humid, and winters were mild, with rare frosts.

The second province was a temperate hill and valley region alternating with flood-plain and

lake-border vegetation. It covered much of the Columbia Plateau and northern Great Basin and extended eastward to the Rocky Mountains. Deciduous hardwoods dominated it. Maple (*Acer*), alder (*Alnus*), birch (*Betula*), hornbeam (*Carpinus*), hickory (*Carya*), redbud (*Cercis*), beech (*Fagus*), sycamore (*Platanus*), poplar (*Populus*), oak (*Quercus*), and willow (*Salix*) were common, as were such conifers as water cypress (*Glyptostrobus*) and swamp cypress (*Taxodium*) which are now confined chiefly to lowland areas of high water tables. Fir (*Abies*), spruce (*Picea*), pine (*Pinus*), and other conifers were present but generally rare for they lived in bordering montane areas. Rainfall over the area probably ranged from 30 to 35 inches, distributed well throughout the year but possibly of biseasonal occurrence. Thus it is inferred that temperatures of this province were warm in summer and cool in winter and snows were common. This region graded southward to interfinger with the Madro-Tertiary Geoflora (Pl. 1).

The third province has montane characteristics. It was dominated chiefly by conifers of the West American Element, such as fir (*Abies*), spruce (*Picea*), pine (*Pinus*), Douglas fir (*Pseudotsuga*), coast redwood (*Sequoia*), and western red cedar (*Thuja*). Their common associates included species of maple (*Acer*), alder (*Alnus*), serviceberry (*Amelanchier*), birch (*Betula*), hornbeam (*Carpinus*), beech (*Fagus*), aspen (*Populus*), cottonwood (*Populus*), oak (*Quercus*), coralberry (*Symphoricarpos*), and elm (*Ulmus*), many of which represent cooler types, compared with species recorded in the more abundant floras of the lowlands. Judging from the requirements of related modern plants, rainfall in these upland areas was probably 35 to 40 inches, distributed more or less equally in summer and winter. Temperatures presumably were cool in summer, but the winters were cold, and snow probably fell regularly.

Madro-Tertiary Geoflora

The Madro-Tertiary Geoflora, which apparently originated in the southwestern part of North America during the early Tertiary, migrated northward with expanding dry climate during the middle and later parts of the period. Madro-Tertiary species were chiefly drought-resistant or drought-deciduous, small- and thick-leafed plants similar to those now living in the semiarid sections of southwestern North America. The California Woodland and Chaparral elements included plants related to the oak and conifer woodland vegetation, and the dense chaparral communities of the summer-dry climate of California. The Sierra Madrean Element comprised woodland species similar to those now in the summer-wet region from Arizona to west-central Texas and southward into Mexico. The Southwestern Chaparral Element included a number of species similar to those living on the margins of woodland vegetation in Arizona, the Edwards Plateau, and Coahuila. The Sinaloan Element represented a drought-deciduous arid subtropical scrub (thorn forest) whose species resemble those now found on the moister warmer margins of the desert in eastern and western Mexico. The regular association of the species of these elements during most of the Tertiary attests to the unity of this geoflora in time and space. Its successive changes in composition and distribution, due chiefly to the trend toward drier climate during the period, provide a basis for age determination (Axelrod, 1938; 1939; 1948).

Plate 1 shows that two major provinces can be discerned in the area of distribution of the Madro-Tertiary Geoflora in Mio-Pliocene time. In the north its species were chiefly members of woodland and chaparral vegetation. Included in the woodland communities were madrone (*Arbutus*), buckthorn (*Bumelia*) cypress (*Cupressus*), ironwood (*Lyonothamnus*), avocado (*Persea*), digger pine and nut pine (*Pinus*), pistacio (*Pistacia*), sycamore (*Platanus*), cottonwood (*Populus*), live oak (*Quercus*), locust (*Robinia*), soapberry (*Sapindus*), laurel (*Umbellularia*), and Mexican chestnut (*Ungnadia*). Chaparral species were distributed in such genera as manzanita (*Arctostaphylos*), ceanothus (*Ceanothus*), mountain mahogany (*Cercocarpus*), bush poppy (*Dendromecon*), silk-tassel bush (*Garrya*), evergreen cherry (*Laurocerasus*), sumac (*Schmaltzia*), and scrub oak (*Quercus*). It is inferred that rainfall over the northern sector ranged from 20 to 25 inches and was

distributed chiefly biseasonally. Temperatures probably were mild to moderately cool. In the more northern temperate areas, as in central to northern Nevada and on the flanks of the Sierra Nevada, the Madro-Tertiary Geoflora occupied exposed slopes and flats adjacent to cooler sites supporting the Arcto-Tertiary Geoflora (Pl. 1).

In the southern sector, woodland and chaparral vegetation were also characteristic in Mio-Pliocene time. But here they were associated with the arid subtropical Sinaloan Element, comprising such thorn-forest genera as acacia (*Acacia*), elephant tree (*Bursera*), paloverde (*Cercidium*), hopbush (*Dodonaea*), palm (*Erythea, Sabal*), kidneywood (*Eysenhardtia*), fig (*Ficus*), lead tree (*Leucanea*), and guamuchil (*Pithecolobium*). As judged from the requirements of related living species which are found regularly in the warmer frost-free sections of Mexico, milder winters probably typified this southern sector where precipitation over the lowlands was on the order of 15 to 20 inches, most of which fell in the warm season.

Neotropical-Tertiary Geoflora

The Neotropical-Tertiary Geoflora included broad-leafed evergreen trees, shrubs, and lianes of living tropical families. Early in the Tertiary, under the influence of a warm mild climate with adequate rain, it ranged northward to Lat. 55° on the Pacific coast and extended far inland. Progressive chilling during the Tertiary is presumed to have restricted it gradually southward to tropical regions. It occupied a narrow coastal strip (Pl. 1) extending from Monterey through the coastal archipelago into southern California during Mio-Pliocene time, when it was dominated by arid subtropical types. Its recorded Mio-Pliocene species are in genera such as custard apple (*Anona*), sea grape (*Coccoloba*), fig (*Ficus*), magnolia (*Magnolia*), lancewood (*Nectandra*), avocado (*Persea*), palm (*Sabal*), and other subtropical types which mingled on bordering slopes with the pine-oak woodland and scrub communities of the Madro-Tertiary Geoflora. Ecological relationships of similar modern plants lead to the inference that rainfall over the coastal area south of San Francisco Bay was near 30 inches yearly, probably distributed chiefly as a summer maximum. Winters presumably were mild and frostless, but summers were probably hot and humid.

Paleobotanical Evidence Regarding Altitude

Mio-Pliocene Floras in West-Central Nevada

To evaluate the elevation of west-central Nevada we must remember that all the Mio-Pliocene floras in that area were relatively similar in floristic composition and represented a mixture of species derived from the Arcto-Tertiary and Madro-Tertiary geofloras (Pl. 1). Many of the 65 woody plants in the Aldrich Station, Chloropagus, Fallon, and Middlegate floras are related to species which dominate living communities on the lower western flanks of the Sierra Nevada, where yellow pine (*Pinus ponderosa*)—white fir (*Abies concolor*) forest adjoins woodland and chaparral communities (Axelrod, *in press*). This local diversity of ecologic conditions is well illustrated by the fact that Sierra redwood (*Sequoiadendron*) is in three of the four fossil floras and occurs today at the lower margins of the Sierran forest, as on the south fork of the Kaweah River, where it mingles with live-oak woodland and chaparral at an elevation of 3500 feet. The zone of overlap between forest and woodland in the present Sierra descends gradually to lower levels northward and at the latitude of the Nevada floras it is found commonly near 2000 to 2500 feet. This altitude therefore approximates that of the floras in west-central Nevada during Mio-Pliocene time. However, the Mio-Pliocene forest of west-central Nevada appears to have lived chiefly on cooler moister slopes a few hundred feet above the sites of deposition (Pl. 1), for forest species are represented by few specimens, compared with the predominant live oaks. Thus, it seems likely that if the average elevation over the lowlands in west-central Nevada during Mio-Pliocene time was greatly in excess of 2500 feet, a larger representation of forest species would be expected in these floras, in much the same way that forest becomes dominant at levels above 2500 feet on the windward slopes of the Sierra Nevada at this latitude today.

East-West Comparisons of Vegetation

Comparisons of the Mio-Pliocene floras of the Sierra Nevada with those in western Nevada show that the latter are most like the floras from the lower foothill belt of the Sierra Nevada, in a climatic as well as a floristic sense. The floras in both areas suggest an annual rainfall of approximately 25 to 30 inches, and in each region they are made up of an intermingling of plants derived from the Arcto-Tertiary and Madro-Tertiary Geofloras (Pl. 1). On the windward slopes of the Sierra Nevada a number of species such as hickory, tupelo, holly, elm, and sweetgum, find their nearest living relatives in the summer-wet sections of the northern continents. Members of these elements had a poorer representation over the lowlands of west-central Nevada at the same time. This suggests that summer rainfall was not only more frequent and in greater amount, but more effective on the coastward slopes where the evaporation rate presumably was lower. The coastward-slope floras also have avocado, magnolia, and other mild-winter types which are rare in the interior, indicating cooler winters in western Nevada compared with the windward Sierran slopes. Nonetheless, their relict occurrence in western Nevada at this time suggests that the region was still under the occasional influence of moderating oceanic climate. These sites may have been opposite low saddles in the range or near the heads of broad valleys up which mild air penetrated to the interior during winter. In any event, the moderate difference between the climate and vegetation on the lower windward flanks of the range and that to leeward in western Nevada must mean that the Nevada plateau was not at a very high elevation; otherwise the climate would have been cooler and the floras would have differed markedly. It is apparent also that the intervening ridge only comprised low hills; otherwise a drier climate would have occurred to the east and resulted in additional differences in vegetation.

Direct evidence of the elevation of the Nevada plateau is afforded by a comparison of the Remington Hill flora from the coastward slope (Condit, 1944a) with the Fallon flora east of the present range (Axelrod, in press). The Remington Hill flora contains coast redwood (*Sequoia*) and species of sycamore, cottonwood, oak (three species), poplar, willow, and madrone which are similar to those found with redwood today in the subhumid parts of its range in the hills north and south of San Francisco Bay. The Fallon flora has a number of plants related to those in the present Sierran forest, including yellow pine, white fir, nutmeg, Sierra redwood (*Sequoiadendron*), Oregon grape, madrone, willow, and cottonwood. Their small representation suggests that they lived on slopes a few hundred feet above the Fallon basin of deposition, which was dominated by live-oak woodland.

Apart from Sierra redwood, most of the typical members of the living Sierran forest occur also in the central Coast Ranges north and south of San Francisco Bay. In this region the lower margin of the Sierran forest is on slopes at elevations generally ranging from 1000 to 1500 feet above the redwood groves, in areas of cooler climate where snows are occasional in winter. The major difference between these environments represents approximately 5°F. in average January minimum temperature, which approximates the temperature contrast between the Remington Hill and Fallon floras from the coastward and leeward sides of the Sierra Nevada during Mio-Pliocene time. This estimate fits the paleoecology suggested by these floras. As judged from the Remington Hill and other near-by Mio-Pliocene floras (*see* Fig. 1), conifer forests of the Sierran type were not prominently developed on the coastward slopes of the Sierra Nevada at that time, which may have been too mild in winter for a forest of this type. But in the interior across the Sierran ridge, winter temperatures were sufficiently low in Mio-Pliocene time to support a Sierran conifer forest of the type commonly found at the margins of woodland and chaparral country. A difference in altitude of approximately 1500 feet between the Remington Hill and Fallon forests is sufficient to explain the temperature difference of 5°F., for the vertical temperature gradient in air averages 3½°F. for each 1000 feet elevation. Since the Fallon flora lived chiefly on slopes above the site of deposition, the difference in altitude between the Fallon and Remington Hill sites must have been less than 1500

feet and probably not much more than 1000 feet. This difference added to the 1000- to 1500-foot altitude which is here postulated for the Remington Hill site results in an estimate of from 2000 to 2500 feet for the lowlands about the Fallon basin of deposition.

TABLE 1.—MIO-PLIOCENE FLORAS FROM THE SIERRA NEVADA, THEIR PROBABLE ALTITUDE, AND THEIR YEARLY RAINFALL, AS INFERRED FROM RELATED LIVING VEGETATION*

Fossil floras	Present position	Probable Mio-Pliocene altitude	Estimated rainfall (inches)
Carson Pass	on crest	near 2500 feet	35–40
Mohawk; Monte Cristo	middle-upper slopes	2000 feet	40–45
Remington Hill; Forest	lower-middle slopes	1000 1500 feet	30–35
Table Mountain	foothills	500 feet	25–30
Valley Springs	low foothills to piedmont	200–300 feet	25–30

* Some of the data presented here for age, altitude, and rainfall differ moderately from those given in an earlier study of these floras (Chaney, Condit, and Axelrod, 1944).

Rainfall Gradient Across the Sierra Nevada

Comparison of the rainfall gradient across the Sierra Nevada during Mio-Pliocene time with that on these slopes today provides an approximation of the relative change in slope and hence in the probable elevation of the range. Table 1 shows that the rainfall gradient on the Sierran slope was not very steep in Mio-Pliocene time. The floras in the lower piedmont near sea level suggest an annual rainfall near 25 to 30 inches which increased gradually to a maximum of from 40 to 45 inches on the middle-upper slopes and thence decreased to 25 inches over the lowlands of western Nevada. This is only a gentle gradient compared with that on these slopes today where rainfall in the Great Valley near sea level at this latitude averages 15 inches and increases to 70 or 80 inches in the wettest parts of the middle-upper slopes, with a decrease at higher levels and a continued decrease to the lee of the range in the Nevada desert, where it is less than 5 inches.

In evaluating these rainfall gradients to arrive at an estimate of altitude, the wettest part of the middle-upper slope is the most sensitive indicator of elevation, for precipitation increases steadily to a maximum at this level and thence decreases. Compared on this basis, the present gradient shows an increase of approximately 60 inches compared with 20 inches for Mio-Pliocene time, which suggests a Mio-Pliocene gradient only a third of the present one. This leads to the inference that the humid middle-upper slopes lay near 2000 feet in Mio-Pliocene time and, as an average, the general summit region was not much more than 1000 feet higher.

Carson Pass Flora

The estimated altitude of the summit area in Mio-Pliocene time is strongly supported by the Carson Pass flora which is on the present crest and lay near the Mio-Pliocene divide. The flora has been collected recently from three localities within 50 feet of the base of the andesite section which rests on a granitic surface of low relief. The flora is chiefly a riparian assemblage of maple (*Acer*), tupelo (*Nyssa*), sycamore (*Platanus*), avocado (*Persea*), poplar (*Populus*), lingnut (*Pterocarya*), and willow (*Salix*). It provides a remarkable contrast to that at the fossil locality today, which is at an elevation of 9100 feet in the arctic-alpine zone of meadowland and fell-fields above the subalpine forest of mountain hemlock and white-bark pine. This flora probably was laid down at not more than 2500 feet and probably was lower. This estimate agrees with the upper limits of sycamore today on the lower western flanks of the range at this latitude, and its fossil relative was one of the dominants of the flora. The absence of conifers in the flora also points to a relatively low altitude. If it had lived much above 2500 feet conifers might have been expected in at least moderate numbers, for they

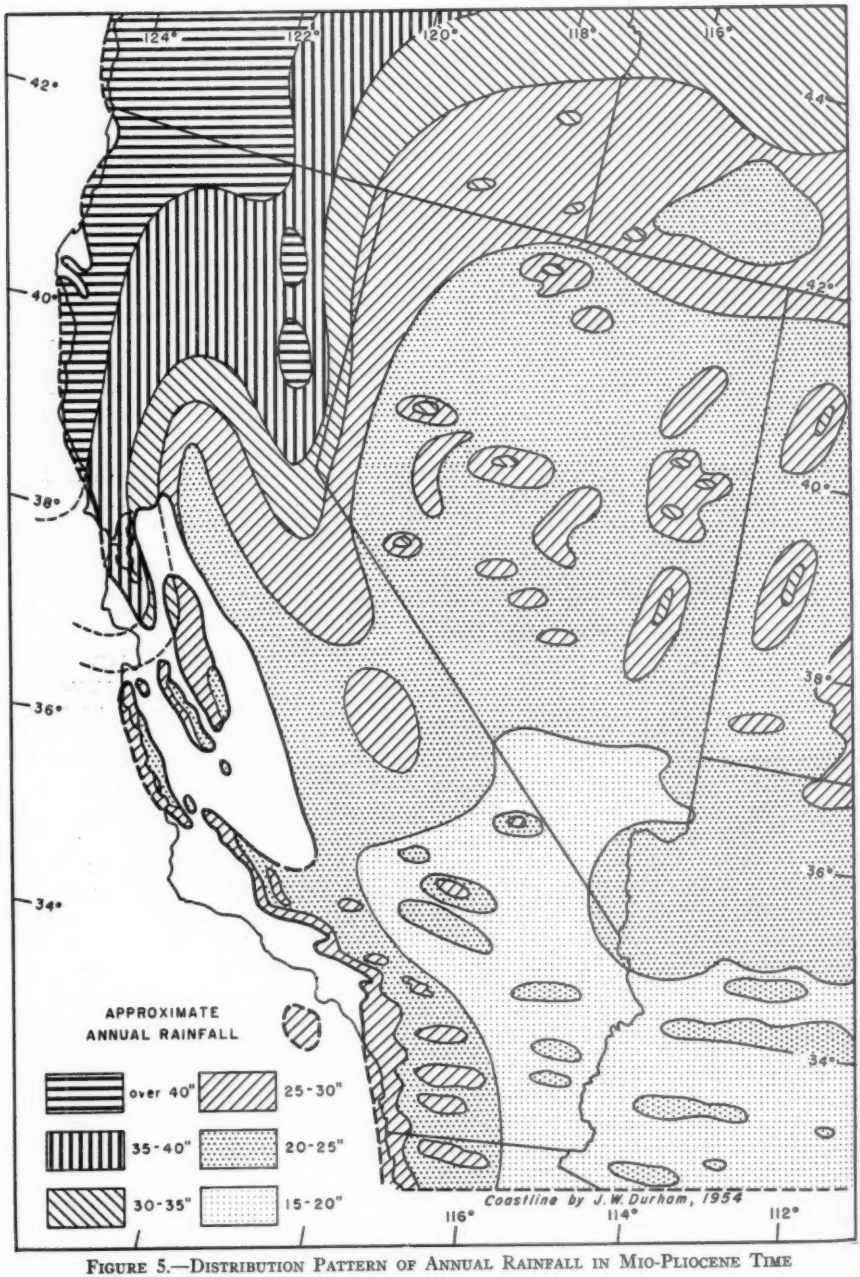

FIGURE 5.—DISTRIBUTION PATTERN OF ANNUAL RAINFALL IN MIO-PLIOCENE TIME
Inferred from data in Plate 1

constitute an important part of all the Nevada floras not far to the east where they blanketed the slopes above the basin of deposition (Pl. 1).

Mio-Pliocene Rainfall

Figure 5, which shows the presumed regional distribution of annual rainfall in Mio-Pliocene time, was drawn on the basis of the rainfall requirements of living vegetation most similar to the Mio-Pliocene floras. It outlines the relation of the Sierran province to west-central Nevada and to the bordering environments at the north and south. In filling out this map for areas where Mio-Pliocene floras are not now known, data from slightly older and younger floras of the region have been used (Fig. 1; Pl. 1). The evidence, although not detailed,[5] indicates that the whole Sierran province was a relatively minor topographic feature at this time. The climatic differences recorded by these contemporaneous Mio-Pliocene floras on both sides of the present Sierra Nevada can be accounted for by assuming that, as an average, the range rose no more than 1000 feet above the Nevada plateau. Judging from the climatic effects of similar low ridges on related vegetation today, it is concluded that if the Sierran summit was much more than 2000 feet above the Nevada lowlands in Mio-Pliocene time, a climate would have existed to leeward which was drier and cooler than that recorded by the Mio-Pliocene floras.

Discussion

These data lead to the inference that in its broadest features the average summit level of the central and northern portion of the Sierran block lay near 3000 feet. In general the late Tertiary volcanics rest on a topography of varied yet generally low relief. In the foothill belt the volcanics were laid down on a gently undulating plain with wide valleys separated by low broad interfluves; locally in this belt old monadnocks of basement rock apparently rose above the general level. In the central strip, the late Tertiary volcanics accumulated on a surface of moderate relief characterized by wide valleys separated by broad rounded ridges rising from 500 to 1000 feet above the valleys. The summit region was slightly more rugged, with the higher hills rising generally from 1000 to 1500 feet above the 2- to 4-mile wide valleys in the passes leading east.[6] In local sections, as in Pyramid Peak range west of Lake Tahoe, a few high hills of basement are believed to have lain sufficiently above the general summit level so that they were not covered by a thick volcanic cover (Lindgren, 1911, p. 37–39). It seems more probable that a somewhat thinner andesite section accumulated there and has since been eroded, possibly because of greater uplift in this local region. A thin cap of Kate Peak andesite occurs a few miles east of the Pyramid Peak range, at a higher altitude. It is on the peak 1 mile southwest of Jobs Peak (Markleeville quadrangle) at an elevation of 10,400 feet, whereas Pyramid Peak is only 10,020 (Carniss Curtis, personal communication, October 1955).

There is some evidence that the range gradually decreased in altitude northward in Mio-Pliocene time, in much the same way that the Sierran block today falls off to lower levels northward where it disappears under the volcanic cover of Mount Lassen. A newly discovered small Mio-Pliocene flora near Juniper Valley (Fig. 1, loc. 19), which lies east of the north end of the Sierra Nevada, appears to be more humid than floras of comparable age in west-central Nevada. If its age is determined correctly, and if larger collections substantiate the general ecologic implications, the barrier at the west was quite low or essentially nonexistent at that time. Thus the summit line of the Sierran block may have decreased northward from the Carson Pass area. In this distance of 100 miles the ridge probably descended to an average level of near 2000 feet at Donner Pass, with an additional drop of 500 to possibly 1000 feet in the next 50 miles to the Lassen area.

Judging from the common occurrence of foreign clasts in the pebble conglomerates in the

[5] Many climatologists have drawn rainfall maps from weather stations that are as scattered geographically as the "stations" recorded here, with results that are as generalized as those reached here.

[6] The author is indebted to Frank S. Hudson who made available a map of the north-central Sierra Nevada, covering parts of the Ione, Placerbille, Colfax, and Truckee quadrangles, which contours the base of the andesite and illustrates the nature of the prevolcanic surface.

basal parts of the andesite section, it is evident that drainage must have been westward across the Sierran ridge (Pl. 1). Paleoecologic data suggest that these seaward-draining valleys probably had elevations not greatly in excess of about 2000 to 2500 feet in the divide area. The floras in the basal part of the volcanic section sediments, such as siliceous shale well-bedded regularly are preserved in lenticular lacustrine water-laid tuffs, and tuffaceous siltstone, which are associated with fluviatile deposits. These relations suggest that small ponds and oxbow lakes occupied the meandering parts of the river valleys which incised the low Sierran ridge.

Paleogeographic Implications

General Statement

The paleoecologic and paleoclimatic data reviewed above for the Mio-Pliocene floras of the Sierra Nevada and west-central Nevada, supplemented by data from floras of similar age in bordering regions to the north and south, appear to complement and reinforce one another with respect to the average altitude of these major topographic features during Mio-Pliocene time. Although the inferred elevations are approximations only, it seems probable that they are within 500 feet of average levels. Since the suggested altitudes have important implications to geologic, climatic, and biologic problems, it is desirable to outline their nature briefly.

Geologic

The preceding estimates of elevation appear to agree with known geologic facts. Figure 6 compares the topographic profile across the region today with that inferred for the interval just before the outbreak of Mio-Pliocene vulcanism in the Sierra Nevada; the volcanoes shown in Nevada are slightly older (*i. e.*, upper Miocene). The profiles were drawn from sea level in the Valley of California through the Lake Tahoe area to the margin of the Carson Sink in western Nevada, a distance of 140 miles. The vertical scale was exaggerated twice to show the topography clearly.

Following vulcanism, diastrophism resulted in major vertical displacement of the range during late Pliocene and early Pleistocene times, and uplift increased southward. The granodiorite-volcanic contact now lies at 7300 feet in the Donner Pass area, but at Carson Pass 50 miles southward the contact is at 9000 feet. Taking 2000 and 2500 feet as the average elevations for the Mio-Pliocene valleys in the Donner and Carson areas at the outbreak of vulcanism, uplift in the Donner area was approximately 5300 feet, compared with 6500 feet at Carson Pass. This increased elevation of the range was caused by at least three factors. In some measure—possibly small and local in extent—it was due to gentle folding (Hudson, 1948; 1951). A good part of it was due to large-scale faulting, which is now expressed as a series of giant treads (basins) and risers (ranges) which, in general, fall off progressively to the east of the main scarp, with the basement lying near sea level in the deeper basins of west-central Nevada (Axelrod, *in press*). The amount of absolute uplift by faulting is difficult to estimate because differential movement on the blocks lying to the east of the crest complicates the picture, as does reverse faulting. The large marginal faults which bound the east-facing scarps of the Sierran block suggest an average maximum figure of approximately 3500 feet for the displacement of the volcanic-basement contact in this section of the range. General regional uplift was also an important element adding to the height of the range following the volcanic episode, as suggested by LeConte (1889) and discussed by Lindgren (1911) and Hudson (1955). It is difficult to estimate the increase in elevation by regional uplift alone, yet if 3500 feet is used as an average maximum uplift that may have been caused by faulting, the remainder may represent the minimum elevation due to regional uplift (Table 2). These figures suggest a gradually southward-rising warp which accounted for a 1200-foot rise of the summit area from Donner Pass southward to the Carson Pass region 50 miles distant in the post-volcanic interval.

To the east, the lowlands of west-central Nevada now lie fully 1500 to 2000 feet above the presumed Mio-Pliocene average level. This increased elevation of the lowlands was caused

PRESENT-DAY PROFILE
Present-day profile from Sacramento across the Sierra Nevada to the west margin of the Carson Sink, Nevada.
L.
Pyramid Pk. Range Tahoe Carson Range Pine Nut Range
2 miles
1
Vert. Scale X2
INFERRED MIO-PLIOCENE PROFILE
Suggested Mio-Pliocene profile across area shown above, as inferred from paleoecologic data.
2 miles
1
Vert. Scale X2
0 10 20 30 40 50 miles
Horizontal Scale

FIGURE 6—TOPOGRAPHIC PROFILE ACROSS THE CENTRAL SIERRA NEVADA TODAY COMPARED WITH INFERRED PROFILE JUST BEFORE MIO-PLIOCENE VULCANISM

in part by basin filling and probably also by regional uplift, for the floors of these basins now lie near sea level (Axelrod, *in press*). All the basins in western Nevada did not originate by faulting, as Nolan (1943, p. 182–184) has implied. In the Middlegate region the sediments clearly lie on an old volcanic terrain, and the basin probably originated by local downwarping. On the northern and southern margins of the Carson Sink, Miocene and younger rocks rest everywhere on pre-Tertiary basement rocks. Both damming by lavas and downwarping may have occurred in basin development here. In the Coal Valley area west of Walker Lake relatively fine upper Miocene sediments lie on crystalline rocks on the east side of the basin, whereas to the west the overlying lower and middle Pliocene section is buttressed against an old scarp. It is apparent that the only generalization that can now be made is that different basins in western Nevada have had different histories, and that each one must be worked out separately.

TABLE 2.—ESTIMATE OF REGIONAL UPLIFT IN THE NORTH-CENTRAL SIERRAN SUMMIT AREA*

	Donner Pass	Carson Pass
1. Present altitude of basement-volcanic contact on Sierran summit in Mio-Pliocene valleys	7300 feet	9000
2. Inferred Mio-Pliocene altitude of basement-volcanic contact of above features	2000	2500
3. Amount of increased elevation due to diastrophism (1 minus 2)	5300	6500
4. Approximate maximum by faulting	3500	3500
5. Approximate minimum by regional uplift (3 minus 4)	1800 feet	3000 feet

* The summit region probably had from 1000 to 1500 feet of relief. The figures in 1 and 2 above represent the elevations of the seaward-draining valleys in the summit area.

Climatic

Factors.—Post - Mio - Pliocene climatic changes in the Far West appear to have been due in part to a secular climatic trend during the Tertiary, and in part to the building up of topographic barriers by vulcanism and diastrophism. Since both were effective concurrently, it is difficult to estimate their relative importance at any one time. The following sketches the nature of the secular trend which culminated in the development of a mediterranean climate and also the probable role that increased relief had on modifying the climate of the region.

Secular effects.—The major climatic trend of the Tertiary involved the chilling of the oceans and the development of drier land areas characterized by higher summer and lower winter temperatures. The broad zonal climates of the early Tertiary developed into the narrow zonal or cellular patterns of today. That the oceans chilled progressively during the Tertiary is evident from the ecological relations of fossil and modern shallow-water marine invertebrates throughout the world. Recent detailed paleoecological studies of the Tertiary faunas of the Pacific coast (Durham, 1950) lead to the inference that at this latitude (40° N.) there was a shift in minimum winter shallow-water temperatures from 25° C. (77° F.) in the Eocene to 11.5° C. (52° F.) at the end of the Tertiary, and the data suggest only a gradual change. The Tertiary Geofloras reflect an increase in the severity of continental climates during the same interval. Most of the dominant subtropical forests representing the Neotropical-Tertiary Geoflora were replaced over middle latitudes by middle Tertiary time by the temperate Arcto-Tertiary Geoflora, except for a narrow coastal strip where tropical relicts are recorded. Woodland, chaparral, and grassland vegetation of the Madro-Tertiary Geoflora expanded widely over the lowlands of the Far

West in the late Tertiary, displacing much of the Arcto-Tertiary Geoflora in its southern sector. Sage and then desert communities dominated the lowlands of this region at the end of the period. Thus the Tertiary Geofloras record a shift from equable climates of ample rain and mild winters to extreme continental climates of low rainfall and wide ranges and extremes of temperature at the end of the period.

As discussed by MacGinitie (1937; 1941), this secular trend toward cooling oceans and more extreme land climates gradually brought into existence a permanent north Pacific high-pressure system during the summer season, particularly following Miocene time. Winds moving from the north Pacific high over the colder ocean would take up relatively less moisture compared with the earlier Tertiary. But more important, the winds blowing onto the land area were being warmed and becoming drier and had lower potentialities for rain in the warm season. By contrast, in the middle and early Tertiary a warm ocean provided moist warm air which rose gradually as it crossed the continent, producing convectional showers during the summer season when the temperature contrast between the ocean and the land was much less pronounced. The temperature gradient between the Eocene tropics which extended to Lat. 55° N. and the cool temperate regions at latitudes above 70° was much less than it is today. The polar front must have been greatly weakened and restricted far to the north in Eocene time, with rare winter cyclonic storms entering middle latitudes where the chief precipitation was of the tropical convectional type in summer time. During middle Tertiary time, with the northern margin of the tropics near Lat. 40° N., conditions essentially transitional to those of the present day may have prevailed. The frequency of winter storms was now probably increased at middle latitudes, with possibly a third to a half of the yearly precipitation falling in the winter season. The decrease in summer rain following Miocene time was particularly marked, as indicated by the successively younger floras in the far West which contain progressively fewer species whose modern correlatives require summer rain. This relationship probably correlates with the continued trend toward chilling oceans and the development of hotter land areas in the summer season, with the resultant reduction in summer rain. To judge from the evidence supplied by the Tertiary marine isotherms (Durham, 1950), the driest continental climate would have occurred in mid-Pliocene time, a conclusion reached earlier on the basis of plant evidence (Axelrod, 1948). It was also during this stage that most of the fossil plants that are indicators of summer rain show a great reduction in numbers, and only a few remained in the coastal region into the Plio-Pleistocene interval. Thus by the end of the Tertiary the secular climatic trend had led to the development of a new climate on the west coast—a mediterranean climate of dry summers and wet winters with precipitation of cyclonic origin chiefly.

Topographic effects.—Vulcanism and diastrophism in late Cenozoic time probably account in large measure for the extreme climates which now prevail over the lowlands of Nevada and on the summits of the Sierra Nevada and the higher basin ranges. There can be no doubt that the secular climatic trend of the Tertiary shifted rainfall distribution from a dominant summer occurrence of convectional showers to a maximum of winter precipitation of cyclonic origin and resulted also in a general lowering of precipitation on the west coast. But it cannot account for the present desert area of Nevada which has only 4 inches of rain. Under present climatic conditions, and with low hills at the site of the Sierra Nevada and the Cascades to northward, the present desert region probably would be sufficiently moist to support a piñon-juniper woodland.

Comparisons of the Mio-Pliocene floras of west-central Nevada with floras of middle Pliocene age in the same region show a change from woodland and forest living under 25 to 30 inches rainfall, to semiarid savanna, grassland, and stream-border vegetation persisting under 15 inches precipitation (Axelrod, 1948; 1950c). This diminished rainfall might be ascribed to the effects of vulcanism which had added approximately 3000 feet of andesite to the Sierran ridge by the close of early Pliocene time. If such a topographic factor were in operation, the floras to the lee of the range should reflect a greater reduction than those of the same age

on the windward slopes of the range or in west-central California (Axelrod, 1944). Since the floras in the three areas indicate approximately the same decrease in precipitation from Mio-Pliocene into middle Pliocene time, vulcanism could not have materially increased the height of the range. This probably was because the crust could not support the volcanics on the Sierra Nevada and adjacent Nevada. To maintain isostatic balance, the basement would have to sink and displace a mass of sima equal to the added volcanics. Since an andesite blanket 3000 feet thick would have to subside approximately four-fifths to come into equilibrium, the surface would be only about 500 feet higher than before vulcanism. This appears to be of the right order of magnitude for the average maximum thickness of the volcanics, for a thicker section would have resulted in a higher ridge, and hence the rain-shadow effect would have been pronounced.[7] Thus the summit of the range may have been at approximately 3500 feet in the middle Pliocene, compared with the 3000 feet inferred for the interval preceding Mio-Pliocene vulcanism. The data suggest, therefore, that the decrease in rainfall recorded by the middle Pliocene floras was due chiefly to secular change and not to a pronounced topographic effect. It was only in post-middle Pliocene time, largely in the late Pliocene to middle Pleistocene interval, that warping and faulting increased the altitude of the Sierran crest line from 5000 to 6000 feet in this north-central section. The effects of warping and faulting increased southward, and 100 to 150 miles away maximum uplift was approximately 7500–9000 feet. This would clearly have accounted for a 10- to 15-inch reduction in precipitation over the Nevada basins, so that widespread desert environment came into existence (Axelrod, 1950c).

The arctic-alpine and subalpine zones now in the summit region of the Sierra Nevada and the higher basin ranges are thus due primarily to a topographic factor alone, and not to a long-term climatic trend. Increasing the height of the Sierran summit 6000 to 8000 feet by vulcanism and diastrophism in late Cenozoic time would have lowered the January temperature from 20° to 28° F. below that which prevailed on the Mio-Pliocene crest. The higher summits had thus attained sufficient height so that snow and ice could persist through the summer. By increasing the altitude of the Nevada lowlands 2000 feet in late Cenozoic time the average January temperature would have decreased approximately 7° F. If this figure is added to the 5° F. difference between the lower Sierran slopes and western Nevada which is postulated for Mio-Pliocene time, the total of 12° F. agrees closely with the present difference in average January temperature between these regions today and hence tends to support the earlier conclusion.

[7] Assuming a thickness of 3000 to 3500 feet (average 3250 feet), the surface would be increased approximately 700 feet; with a load of from 4500 to 5000 feet (average 4750 feet) the surface would rise approximately 1000 feet.

Biologic

These climatic and topographic changes over the Sierra Nevada and west-central Nevada during the later Tertiary opened new ecologic zones for plants and animals. Among the new sites in the uplands of both regions were montane summits characterized by (a) arctic-alpine meadows and fell-fields, (b) a zone of subalpine forest, and, in addition, in the Sierra Nevada, (c) a moist-cold zone of fir forest. Over the lowlands, successively drier and more extreme environments came into existence, as is recorded by the gradual displacement of woodland and forest by steppe (sage, grass) and desert vegetation (Axelrod, 1950c).

The advent of a mediterranean climate at the close of the Pliocene had an important effect on plants in these new zones, as well as in the older forest, woodland, chaparral, and grassland communities bordering them. Many types which had lived under summer rain were eliminated as effective summer showers disappeared, and communities were opened for types tolerant of dry summers in environments ranging from the humid west slope of the Sierra Nevada across the new subalpine tract and into the desert at the east. In herbaceous groups particularly, hybridization and other genetic changes may have given rise to many new types which could, at least marginally at the outset, invade these new zones in a novel climate. Microgeographic isolation in these areas of

abrupt and diverse changes in climate and relief would permit the existence of divergent adaptive types in proximity. Further hybridization would favor the development and spread of types more strongly adapted to the extreme environments. Fluctuating glacial and pluvial climates probably intensified the amount of hybridization, for rapid climatic fluctuations—such as four glacial advances and retreats in the Wisconsin—must have continually altered the spatial relations of populations. Continued crossing of divergent, but rather freely recombining, genotypes would establish new variants even more narrowly adapted to these extreme rapidly fluctuating environments. In this way scores of highly adapted types may have arisen quickly, compared with rates of evolution in related groups occupying more uniform areas.

Figure 7 summarizes the shifting relations of the major adaptive plant zones in this region during the later Cenozoic; they might also be termed subzones of a montane zone. The invasion of a new zone by a group of organisms clearly represents a basic evolutionary event which involves secular change of the zone in a physical as well as a biological sense (Simpson, 1953, Chap. 7). As suggested by Simpson's discussion, the broader zones express a more generalized adaptive type, whereas the narrower ones (desert and arctic-alpine) are more specialized. The modal curves shown in the figure suggest the direction and intensity of selection, with the asymmetry implying a linear component imposed by shifting climatic factors in time. Although the actual picture was far more complex than can be shown (Fig. 7), the over-all pattern nonetheless clearly suggests the expanding opportunities for life in this region during the later Cenozoic.

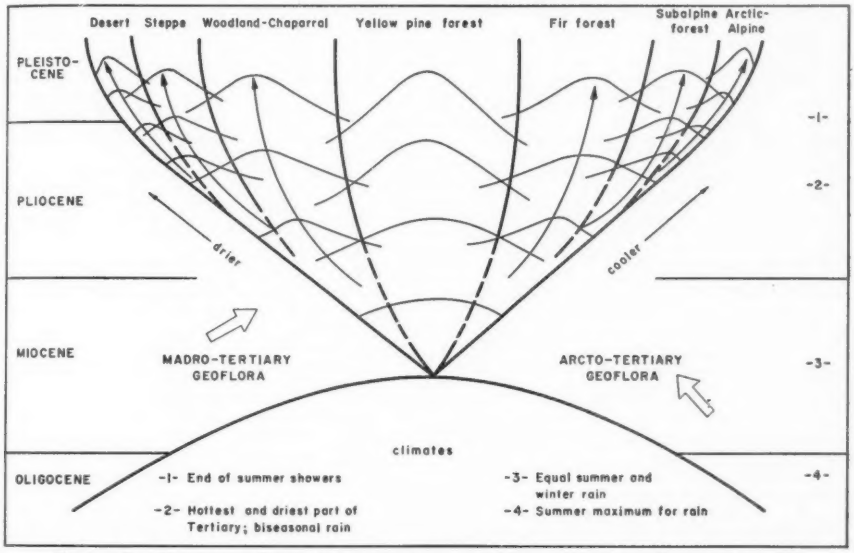

FIGURE 7.—EXPANDING ZONAL RELATIONS IN THE REGION DURING THE LATER CENOZOIC

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MILITARY GEOLOGY IN THE UNITED STATES SECTOR OF THE EUROPEAN THEATER OF OPERATIONS DURING WORLD WAR II

BY CLIFFORD A. KAYE

ABSTRACT

Geology, which was of far-reaching importance on the Western Front of World War I, played a less spectacular role during World War II in so far as the United States armies in Europe were concerned. The U. S. Army in the European Theater of Operations (ETO) used geologists in two capacities: (1) to make staff studies at the level of Theater Headquarters, and (2) as officer personnel in a water-supply unit. In the first category, only one group of seven geologists was employed, forming part of the Information Section, Intelligence Division, Office of the Chief Engineer, ETO. The products of this group consisted largely of regional and localized terrain (trafficability) studies that ranged geographically from the Normandy invasion beaches to Czechoslovakia. Problems of water supply, sources of road material, and many other questions of a geologic nature also arose. French geologists collaborated closely with the work of the section. As the campaign progressed, the Military Geology Unit of the U. S. Geological Survey made important contributions to the geologic intelligence of Germany.

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INTRODUCTION

During the last decade the thought must have crossed the minds of many that the European campaign of World War II produced no report on the role of geology and geologists comparable to Brooks' (1920) account of American geologic work during World War I, or the fine volume on the geologic work of the Royal Engineers on the Western Front (Anonymous, 1922). In the absence of any published description of such work, the question must have been raised whether there was any military geology at all within army echelons in the American sector of Europe during World War II. The answer is in the affirmative, although this work was organized on a smaller scale than in World War I. To understand this one must take into account the differences in the nature of the military operations of the two wars: static trench warfare in the first, and highly mobile and mechanized warfare in the second (particularly during the post-D-day phase). This shifted the emphasis from military geology of a more stratigraphic nature and of very great tactical value in the first conflict to topographic geology primarily of a strategic value in the second, and from field geology to a somewhat less dangerous office geology.

Because some account should be given of this

work, and in order to supplement Hunt's (1950) description of the work done by the Military Geology Unit of the U. S. Geological Survey in all theaters of operation during World War II, this paper is written from the personal recollections of the writer.

Acknowledgments

Acknowledgment is due the following for helping to refresh the writer's memory: Alice S. Allen, Manuel G. Bonilla, Edwin B. Eckel, Maxim Elias, Lyman Huff, Frank Spencer, and Stewart Wallace. The writer wishes to thank Frederick Betz, Jr., Military Geology Branch, U. S. Geological Survey for critically reviewing the manuscript.

Differences between World War I and World War II

The geologic work on the Western Front of World War I has been admirably described for the U. S. Army by A. H. Brooks (1920) and for the British armies in a volume from the many-volumed report on the works of the Royal Engineers (Anonymous, 1922). This geologic activity was primarily concerned with ground-water resources, suitability of terrain for entrenchments and underground mining (military mining), and sources of road-metal and other construction materials.

Ground water was essential for the water supply of many of the front-line areas of World War I and for military hospitals and depots where military concentrations exceeded the capacities of surface-water resources. It also played a decisive role in entrenchment and military mining. The trench warfare that characterized the First War required much geologic advice. Entrenching was possible only where surface materials were easily excavated by pick and shovel. This ruled out as combat zones, or should have, (*see* Brooks, 1920, p. 87) areas where hard rock lay close to the surface, and the military planners needed geologic advice in these matters. Moreover, trenches, dugouts, and other subsurface fieldworks had to be free draining if they were not to turn rapidly into canals and pools. The geologist was therefore required to predict whether entrenchments of the required depth would be floored in sediments of high or low permeability and whether or not they would be above or below the water table. This called for extensive knowledge of local stratigraphy and for the preparation of detailed geologic maps.

Underground mining was one of the special-attack techniques used in World War I and consisted of tunneling under the enemy, rather than approaching him on the surface or from the air. The British Report (Anonymous, 1922) is especially rich in descriptions of the use of geology in this type of warfare. The geologist served to guide the tunneling to the confines of favorable geologic horizons and, where possible, to levels above the water table. This demanded close geologic supervision of tunneling operations, although, as the reports of World War I pointed out, there was far from universal awareness of this principle, at a cost of much wasted effort.

Lastly, rock for road construction and railway ballast was much in demand, particularly in areas of clayey soil where the heavy military traffic converted country roads into quagmires. Geologists advised on quarry sites, as well as on other mineral matters that came up during the campaign.

The geologists of World War I devoted much time to making test borings and to geologic field studies. This field activity was possible because of the slow-moving or stabilized nature of the front. In the Second World War, particularly in its second, or post-D-Day, phase, the front often moved with the speed of tanks, or faster, and there was little time for field studies of test borings for front-line requirements. Except for shallow foxholes, there were few entrenchments, and military mining was not resorted to in World War II in the post-D-day campaign.[1] Superior water-purifying equipment used in World War II made surface-water supplies adequate for most military needs.

The geologists' role in the Allied armies in World War II was therefore somewhat different from what it had been 25 years earlier, although it was probably no less important. World War I

[1] The underground defenses of the World War II were unparalleled in extent, but most were built before the entrance in combat of American ground forces.

geologists had been able to offer the combat effort information on many aspects of field geology, but emphasis in World War II was placed on the office preparation of terrain appreciations for planning purposes; the geologist's most essential assets were the ability to read topographic maps and aerial photographs, to extract information from geologic or soils maps, and to organize and present graphically vast amounts of diverse data. In the last months of the campaign, the rapidity of the movement of the front was such that it is no exaggeration to say that often all that seemed necessary was a good road map; the direction of several thrusts was primarily predicated on the existence of good roads.

It is noteworthy, however, that both German and Russian armies maintained large military-geology staffs during the course of World War II. Besides fulfilling functions similar to those of the geological units of the Allied armies, German and Russian military geologists were attached to field units and provided day-to-day advice on a tactical level. This fact alone suggests that the Allied forces did not make full use of the benefits that military geology had to offer. Undoubtedly, if the post-D-day phase of the war had lasted longer, the need for better-organized military-geology units within the British and American armies would have become strongly manifest.

Early Planning

As far as the writer has been able to learn, early planning for the continental invasion of Europe from British bases included no work by U. S. Army geologists. Instead, geologic aspects of these preparatory studies were taken care of by the Inter-Service Topographic Department (ISTD) of the British, at Oxford, and by the Military Geology Unit of the U. S. Geological Survey, in Washington, D. C. For several years before the D-day invasion, the geographic and geologic sections of the British group issued many reports on areas that later fell within the U. S. sector and proved very useful to the U. S. Army geologists in preparing more-detailed studies for the continental campaign. Among these reports were small-scale terrain studies of France, detailed studies of the major rivers of western Europe, maps showing quarry sites in France, and maps of possible airfield sites in France. The Military Geology Unit of the U. S. Geological Survey made a series of small-scale terrain maps of all the coastal countries of western and northern Europe and a terrain-appreciation folio of France for General staff planning.

In addition to this work, the Geographical Section, General Staff (GSGS) of the British Army, copied and printed in full color, for general distribution, all the 1:80,000-scale geologic maps of the Carte Géologique de la France. The excellent geographic studies of the countries of western Europe made by the British Naval Intelligence Division (NID handbooks) at their Cambridge subcenter provided background material of inestimable value for the U. S. military geologists.

Contributions in military geology by the British groups to the American section were minor following the assumption of full-scale staff work in the theater by U. S. forces in the spring of 1944.

Organization

The geologic staff of the British Forces on the Western Front in World War I consisted of 5 geologists at the conclusion of hostilities. All these men were assigned to general headquarters (British); 1 was on the staff of the Chief Engineer, and 4 were in the Inspector of Mines office, also under the Chief Engineer. In addition, "Among the officers of the mining troops [were] many mining engineers who had been trained as geologists and were well qualified to extend the observations and deductions of the headquarters staff" (Anonymous, 1922, p. 94). In contrast to this, the British 21st Army group of the last phase of World War II had the services of only 2 geologists (King, 1951).

In the American Expeditionary Force of World War I a total of 18 geologic officers had been authorized by mid-1918, although only 9 positions had been filled by the time the armistice was signed later that year (Brooks, 1920). Of these 9 geologists, 5 were at general headquarters (U. S.), 2 were with the First Army, 1 was with the Second Army, and 1 was with the water-supply section. The authorized quota of 18 geologists was to have been distributed as follows: 6 at general headquarters, 5 with each

of the 2 armies, and 2 in the communication zone (supply service).

In World War II, the use of geologists within the U. S. armed services in the European Theater of Operations (ETO)[2] was limited to a company officer of a water-supply unit and a group within the Information Section of the Intelligence Division of the Chief Engineer. The latter organization was of a tenuous nature, and the geologist's position was not explicitly stated in the table of organization. The water-supply unit was a company that operated with the Seventh Army in southeastern and eastern France. Its primary mission was the drilling of water wells and the operation of major water points. Its company commander was an experienced ground-water geologist.

Whether geologists were used in the headquarters staffs of any of the armies, corps, or divisions to prepare terrain reports is not known to the writer. While this possibility exists, the writer's acquaintance with the terrain-intelligence groups of several of the higher-echelon staffs did not bring him in contact with geologists, and he did not hear of any geologists. Most terrain-intelligence staffs were made up of men from other professions. The terrain-intelligence staff of Supreme Headquarters (SHAEF), for example, consisted of a geographer, an architect, and a schoolteacher.

Information Section, Intelligence Division, Office of Chief Engineer

The Information Section of the G-2 staff of the Chief Engineer was primarily concerned with collecting and disseminating information on enemy engineering equipment (including land mines), information on roads, bridges, railroads, and highways, and information on all aspects of terrain. Geologists were assigned to this section to provide the information on terrain. During the planning stages of the Normandy invasion, the commanding officer asked for the transfer to his section of personnel from within the theater who had professional qualifications as geologists. By the end of hostilities, seven men with varying degrees of geologic training had served in the section. At no time were all seven engaged in what might be called military geology, and generally no more than three or four were actively pursuing these studies. In addition to the geologists, a forester and a student of architecture worked on terrain studies.

There was no rigid chain of command in the group. All geologists were transferred to the Information Section without change in grade, although most of them received a grade raise during their service. Rank was no indication of importance in the military-geology program, and one of the most valuable men was a private during most of the campaign.

Clerical support consisted of typists, draftsmen, and translators. The latter worked on French and German documents, some of which were reproduced and distributed directly to the armies, and others used by report writers.

The Information Section was at first stationed on Upper Brook Street, near Grosvenor Square, London. In August 1944, the section moved to the continent; and from September 1944, to the end of hostilities the section was stationed on Avenue Kléber, Paris, within reach of libraries and the Sorbonne, which were of invaluable aid to the work of the group. During this post-D-day period, the Office of the Chief Engineer was part of the headquarters staff of the Communication Zone.

Although the products of the military geology group in the Information Section were varied, the major occupation was making cross-country movement maps, or terrain appreciation maps, and reports to accompany these maps. The earliest products were maps of the Normandy invasion beaches which showed the main terrain units and the major beach exits. An official observer for the section accompanied the D-day invasion of Normandy. The first military-geology map made by the group, on a regional basis, was an attempt to recast the legends of several of the 1:80,000-scale geologic maps of the Carte Géologique de la France into terms significant to military operations. This early attempt at a military geologic map proved of academic interest only, for military operations

[2] ETO included the operational area of France, Germany, and northern Europe, including Great Britain. It did not include Italy, which was in the Mediterranean Theater of Operations. Planning studies for the Allied invasion of southern France were carried on in the Mediterranean Theater.

passed over the area before the maps could be printed and distributed. These were the only regional maps of a geologic nature made of France. The rapid movement of the Allied armies across northern France after the breakthrough at St. Lô left the terrain studies of France far behind.

FIGURE 1.—MAP OF GERMANY SHOWING MAJOR TERRAIN CLASSIFICATION AND ADVANCE OF U. S. ARMIES IN WORLD WAR II

In August, 1944, the Information Section was asked to prepare terrain reports for all the anticipated U. S. sector of Germany (Fig. 1). This was the first request for information on Germany and preceded the entry of U. S. forces onto German soil by only a few months. Four reports were prepared for each of four map sheets on a scale of 1:250,000. The maps that accompanied the reports were cross-coontry movement maps and classified the terrain into five categories: difficult going, moderately difficult going, moderately easy going, easy going, and swamps and poorly drained land. The criteria used were steepness of prevailing slope and the presence or absence of forest cover. Except for the poorly drained category, soil characteristics were not taken into account. In addition to the slope appraisal, important corridors (passes and gaps) and obstacles (escarpments and mountain fronts) were indicated. The source material used for these studies was German topographic maps (scales 1:25,000 and 1:100,000) aerial photography, and miscellaneous books, pamphlets, and maps which pertained to the regions. A descriptive text which outlined, among other things, important soils factors not indicated on the map, a set of representative ground photographs, and, in one region, a perspective ink drawing accompanied the trafficability map. Descriptions of the rivers and towns were prepared by the nongeologic personnel of the group.

Trafficability studies by the Information Sec-

tion of the Lower Rhine Plain in preparation for the 1945 spring offensive provided one of the first indications that, in general, there would be no "going" problem. The Lower Rhine Plain report consisted of trafficability maps on a scale of 1:100,000, with a tabular text printed on the reverse explaining the characteristics of the map units. The morassy condition anticipated by the military planners for this area in late winter and early spring did not exist, as borne out by a study of the excellent 1:25,000 geologic maps of the Preussische Geologische Landesanstalt.

Photographic interpretation was widely used. In one of the early studies aerial photographs were used for the production of a map showing the distribution of sand, mud, and shallow bedrock on the tidal flats of Quiberon Bay in Brittany. This was part of a study to determine the suitability of the area for the construction of a major port, a plan which was abandoned with the capture of Cherbourg. Subtle differences in color, texture, and pattern of drainage lines were used as guides. A method which made use of the shape of bomb craters was devised for interpreting the texture of unconsolidated sediment from aerial photographs. This technique first came into use when aerial photographs of the swamps west of Nantes showed that bomb craters in peat had a different form from those near by in nonpeaty marsh soils. From a study of areas of known soil conditions, it was established that highly compressible, highly cohesive, and noncohesive earth materials could be readily differentiated on photographs by the shapes of bomb craters. The essential criteria were the profile of the crater proper, the profile of the rim, and the texture of the aureole or apron of crater material that fell beyond the rim. Thus, cohesive soils and shallow bedrock tended to have U-shaped craters, fairly steep rims, and aureoles that contained lumps; noncohesive sands and gravel had V-shaped craters, moderate rims, and smooth-textured aureoles without lumps; compressible soils (muck and peat) had narrow craters, large swollen rims, and small aureoles. The technique was valuable in areas where soils information was lacking and where reconnaissance photographs recorded bomb craters of fairly recent origin. The occurrence of bomb craters in the field of France and Germany was not unusual because of widespread tactical bombing of roads, railroads, and enemy emplacements. The bomb-crater method was applied with advantage in a study of an area in the vicinity of Frankfurt which was under consideration for an airborne drop.

Considerable library research went into a study of the foundations of the Rhine bridges, all of which had been destroyed during the campaign. The Army engineers planned to replace many of these bridges with semipermanent structures as soon as the campaign permitted, and a knowledge of the pier foundations was essential to determine whether end-bearing or friction piles would be required, or whether it would be infeasible to use piles because of shallow bedrock. These questions were answered for each bridge by reference to the German engineering and geologic literature which was available in Paris.

The greatest failure of terrain intelligence in the ETO was the realization in early April 1945, exactly 1 month before the end of hostilities, that no military terrain studies of the Bavarian and Austrian Alps existed on a scale large enough to be of real value. With the anticipation of an imminent withdrawal of the remnants of the German armies to this area—then referred to as the German National Redoubt—four people (one of whom was detailed from the Information Section) tried for 2 weeks to simplify into military terms the vast complex of the Alps. Only one was a geologist, and only one had ever been in the Alps. The resulting maps, on a scale of 1:100,000, were not put to the test, since the German National Redoubt never materialized.

With the end of hostilities, all geologic studies of the Information Section came abruptly to an end.

Military Geology Unit, U. S. Geological Survey

The Military Geology Unit of the U. S. Geological Survey (Hunt, 1950) made a substantial contribution to the ETO early in 1945. In late December, 1944, three members of the Geological Survey arrived in Paris as civilians on a mission not directly connected with current military operations. When they saw the limited facilities of the military-geology group in the Information Section, they offered to make available to the theater the large and experienced

organization of the U. S. Geological Survey in Washington, D. C. Accordingly, a request for trafficability maps covering all the U. S. sector of Germany north of the Alps on a scale of 1:100,000 was transmitted to Washington. (*See* Hunt, 1950, p. 309.) In a very short while (1 week for the first north-south tier) the negatives of the overlay plates of these maps arrived at the ETO, were printed in Paris in the printing plant of the French weekly *L'Illustration* (which had been requisitioned by the Intelligence Division, Office of Chief Engineer, for a map-printing plant), and were widely distributed among all U. S. Army echelons. The maps used both soil and slope factors and had a tabular text printed on the reverse. This series of maps was sorely needed and filled a serious gap in the terrain intelligence of the theater.

Two members of the Military Geology Unit visited the Information Section in the spring of 1945 to collect criticism and comment on the 1:100,000-scale trafficability maps. During their stay they also contributed a report on the ground-water resources of Berlin.

In the final months of the campaign and after the surrender, four members of the U. S. Geological Survey investigated and reported on the substantial work of German geologists, in and out of the military service, and appraised their contributions to the strong German resistance of the last years of the war.

Appraisal

Although geologic factors were important in military operations in the ETO, the war was ended with relatively little awareness of this fact by the army. Military geology contributed to high-level staff work; but at the operational level, with the exception of a water-supply unit, comparatively little conscious use was made of the earth sciences. Unlike the campaign in the Pacific theater, the nature of the campaign in the ETO reduced geology as a tactical factor to a minimum. The dense road net of western Europe and the great mobility of the front during much of the campaign necessitated a minimum of cross-country deployment when contrasted with the jungle operations of some of the Pacific islands. The major corridors of Germany were again followed (Fig. 1). The relatively large tracts of open farm land, the temperate climate, and a system of agriculture similar to our own, made for a familiarity with the European landscape that our troops lacked in the Pacific. There is no doubt, however, that greater awareness of the especial usefulness of geologists in supplying intelligence on important terrain factors, as well as on the many matters involving minerals and earth materials that arise during a major campaign, would have resulted in the formation of a larger and better-organized military-geology unit or units; in this regard the lessons of World War I had been lost to World War II.

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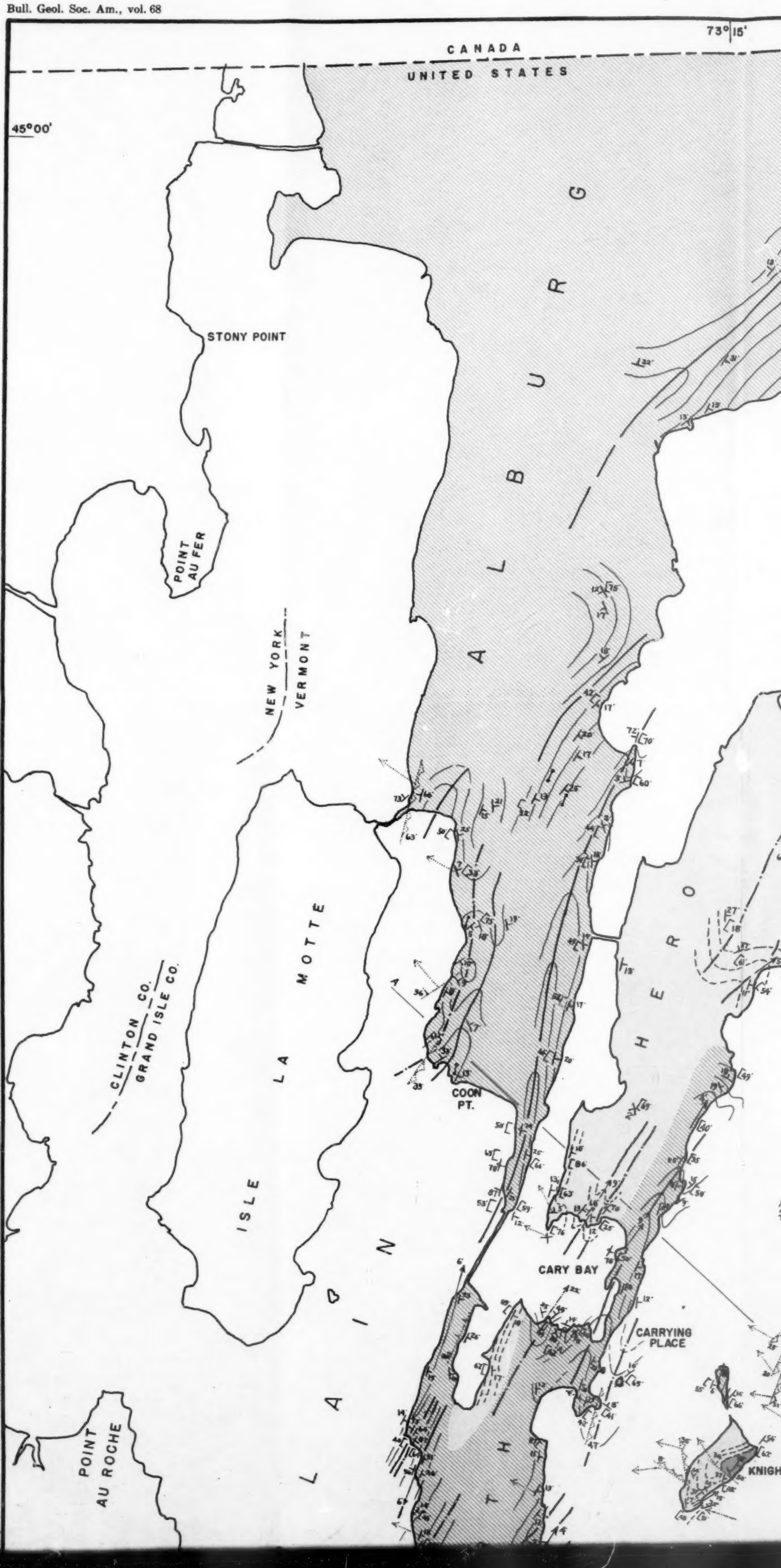
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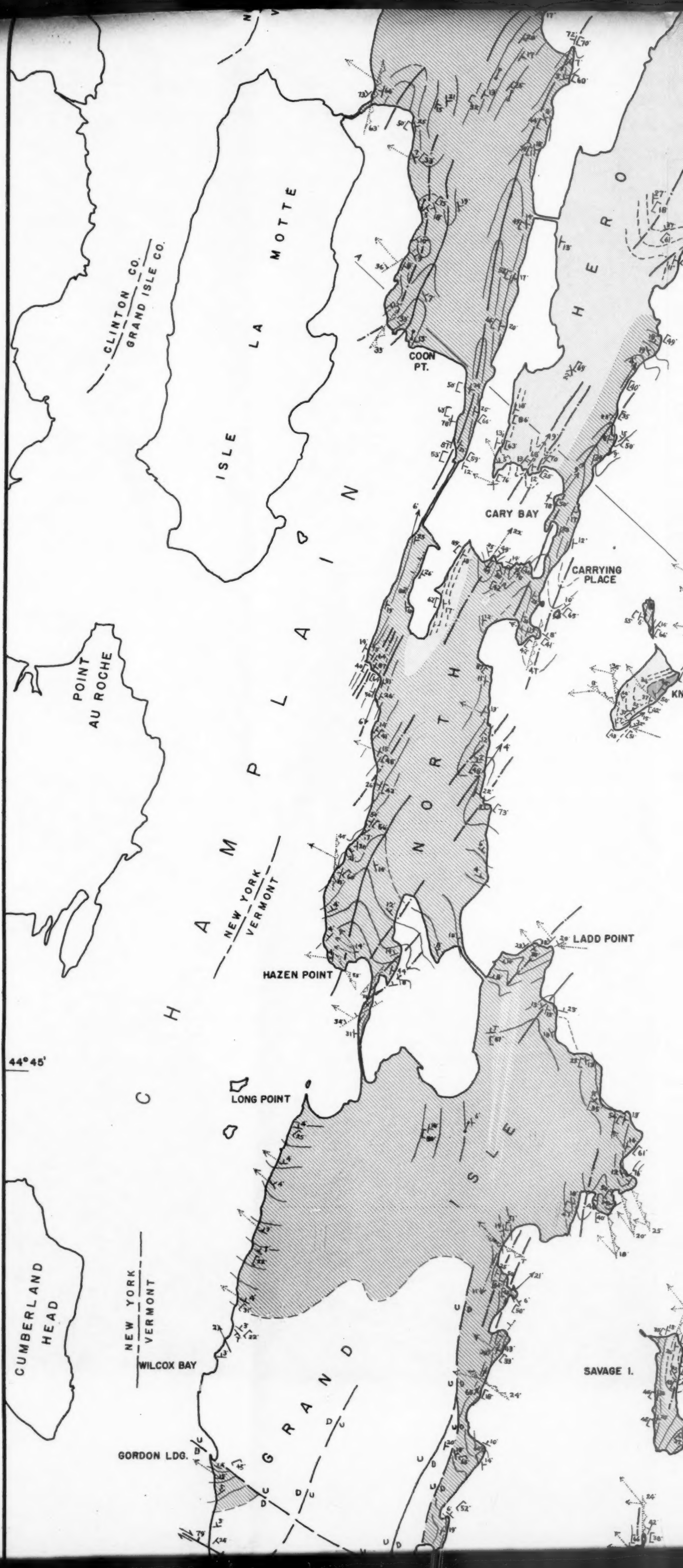
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LEGEND

ORDOVICIAN

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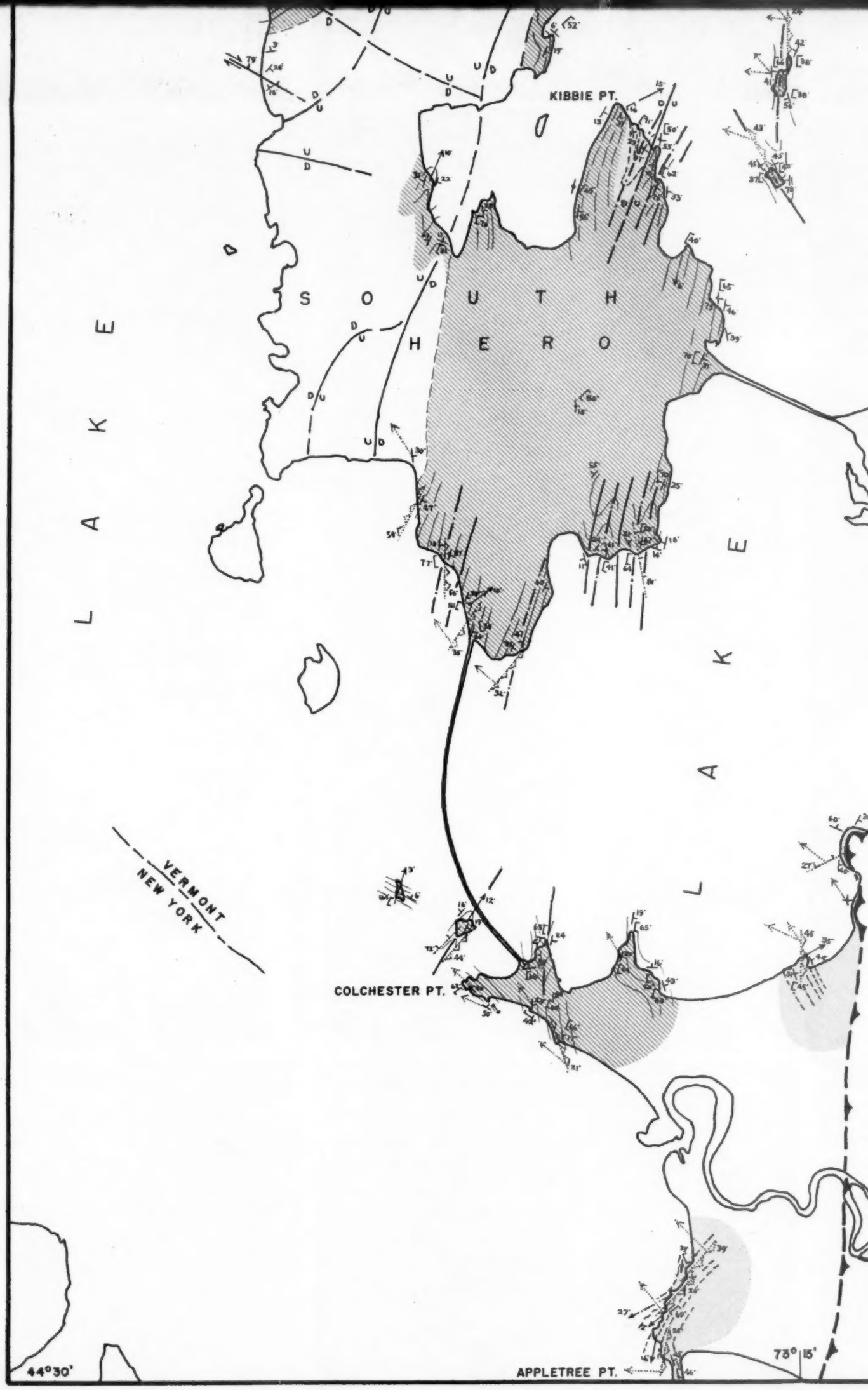
- Iberville Formation (Utica) showing inferred bedding trends
- Stony Point Fm. (Canajoharie) showing inferred bedding trends



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MINOR STRUCTURES HAVE BEEN OMITTED.

HOZ. SCALE IN MILES

VERTICAL SCALE OF PROFILE IS EXAGGERATED

STRUCTURE MAP AND CROSS SECTION OF THE AUTOC

showing inferred bedding trends

- Strike & dip of beds
- Strike & dip of overturned beds
- Strike of vertical beds
- Horizontal beds
- Strike & dip of cleavage
- Strike of vertical cleavage

Fold axes:

- Anticlines
- plunge
- Synclines
- plunge

Faults:

- Dip indeterminable
- Strike & dip of major thrusts
- Strike & dip of minor thrusts
- Strike & dip of minor normal faults
- Strike & dip of strike-slip faults

Bearing of slickensides:

- On fault surfaces
- On bedding surfaces

SCALE

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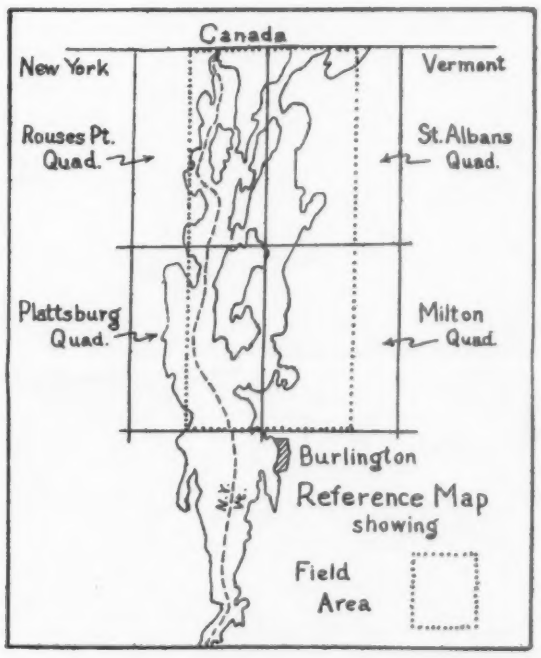

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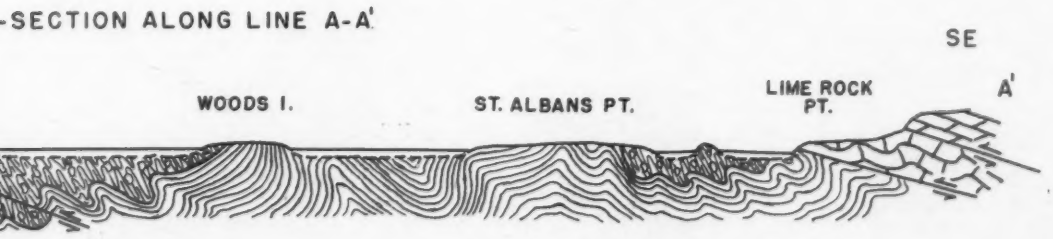

SECTION.

LEGEND

ORDO.	TRENTON	HATHAWAY FM.	
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OCHTHONOUS SHALES IN NORTHWESTERN VERMONT

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 55-94, 3 FIGS., 5 PLS. JANUARY 1957

ORDOVICIAN SHALES AND SUBMARINE SLIDE BRECCIAS OF NORTHERN CHAMPLAIN VALLEY IN VERMONT

BY DAVID HAWLEY

ABSTRACT

The fossiliferous, relatively pure Isle la Motte formation and Larrabee and Shoreham members of the Glen Falls formation (lower Trentonian) indicate deposition in shallow, comparatively clear seas. Upward through overlying Cumberland Head and Stony Point formations there is an increasing proportion of argillaceous limestone and calcareous shale. Small-scale current bedding in the Stony Point formation indicates currents from southwest. Next younger Iberville formation (upper Trentonian), 1000 to 2000 feet of rhythmically interbedded clay shale and graded dolomitic siltstone, contains occasional 5- to 10-inch coarser-graded beds with shale fragments, intrastratal crumpling, flow (load) cast, and slump structures. A deeper, muddier bottom subject to repeated turbidity currents is inferred. Uniformly oriented, small-scale current bedding indicates a northeastern source. Lying on Iberville is the Hathaway formation (new name), composed of argillite and bedded radiolarian chert with blocky inclusions (up to 50 feet long) of limestone, dolomite, dolomitic quartz siltstone, sandstone, coarse graywacke, and chert. It is chaotically deformed with extensive flowage and isoclinal folding perpendicular to regional trends, unrelated to tectonic structure of older formations in the area or to its own cleavage. Submarine slides are inferred, possibly from a thrust fault scarp to the east or northeast.

This sequence of rocks is complexly folded and sheared, with greater intensity near the major "Logan's line" thrusts bounding the area on the east.

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Introduction

The central lowland of the Champlain Valley, northwestern Vermont and northeastern New York, is underlain by Upper Cambrian and Ordovician sedimentary rocks, bordered on the west by the Adirondack Mountains of Precambrian crystalline rock upon which Upper Cambrian sandstone lies unconformably, and against which sedimentary rocks have been dropped along normal faults. The lowland is bordered on the east by low-angle thrust faults on which massive dolomite, quartzite, and limestone, as old as Lower Cambrian, from the east over-rode weaker Ordovician shale and limestone. The westernmost thrusts, the Highgate Springs thrust in the north half of the area, and the overlapping Champlain thrust in the south, trace an irregular line a few feet to 3½ miles inland from the east shore of Lake Champlain.

The shales, the youngest rocks of the autochthonous lowland sequence, were mapped from the Canadian border southward for 40 miles to the vicinity of Burlington, Vermont. Their outcrop includes most islands of Lake Champlain in Vermont, and mainland between the thrusts and the lake. Although exposures on almost continuous shore-line bluffs were excellent, there were few outcrops inland because of glacial cover and low resistance of shales to weathering. Fossils are rare in the older shales and absent in the younger. Lithologic sequence was established almost entirely on structural criteria.

West of the major thrusts and east of the New York-Vermont State line, studies of rocks older than the Cumberland Head have been made by Marshall Kay, and his former students Philip Oxley, Richard W. Rush, and Richard Riggs. Areas underlain by these older formations, western South Hero and Isle la Motte, were omitted from this study.

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Previous Work

Hitchcock delineated principally by lithologic sequence two formations conformably overlying the Trenton of the northern Champlain lowland (Hitchcock *et al.*, 1861, v. 1, p. 301). He correlated the lower black calcareous shale with the Utica shale and the upper "clay slate," with the "Hudson River group" of the New

York State Survey[1] (p. 308). As part of the "Hudson River group" Hitchcock described "beds of silicious slate, . . . a mass of flint or quartz rock" on St. Albans Point, and brecciated limestones with fragments derived from older limestones. He failed to recognize the major thrusts and thought this group continued upward into limestone and dolomite of the Highgate Springs and Rosenberg slices.

The calcareous shales had been correlated faunally with shale at the Rural Cemetery at Albany, New York; when its age was changed from Utica to middle Trentonian (Canajoharie), Ruedemann made a similar change in age of the shales of Champlain Valley (1921a, p. 112–115; 1921b, p. 92–93). Thus, the youngest sedimentary beds of the northern Champlain lowland were considered equivalent to the Canajoharie, and workers lost sight of the noncalcareous shale, chert, and sedimentary breccia originally recorded by Hitchcock.

Stratigraphy

General Statement

Table 1 summarizes the present classification of autochthonous Trentonian formations in northern Champlain Valley.

The sedimentary formations underlying the autochthonous lower Trentonian limestones in this region begin with the Upper Cambrian Potsdam sandstone which rests unconformably on Precambrian crystalline rocks of the Adirondack dome. This, and overlying Canadian (Beekmantownian), Chazyan, and earlier Trentonian dolomites and limestones have been discussed by Brainerd (1891), Brainerd and Seely (1888; 1890; 1896), White (1899), Perkins (1902; 1904; 1921), Cushing (1905), Gordon (1923), Kay (1937), and Buddington and Whitcomb (1941).

A detailed study of the overlying shales seems to have been delayed by their great thickness and apparent lithologic uniformity, and by the complexity of their deformation. Mappable shale units, measurable in hundreds of feet, contrast with much thinner distinct formations below them. The shales, caught between the soles of major thrusts and underlying massive limestone and dolomite, were intensely folded, and sheared. Because of the thickness of the shale units and paucity of their fauna, there are few places where displacement on faults is great enough to juxtapose different units and thus afford a measure of the movement. Deformation has limited the number of measurable sections, but every section of significant thickness was measured and related to other such sections by study of their relative structural positions. Thus, minimum thicknesses for the shale formations were determined.

Lithologic terms are used in this study according to the following definitions.

Shale designates fine-grained (silt or clay size) argillaceous rocks having bedding fissility. Shale that effervesces with dilute hydrochloric acid (1:10) is calcareous shale. Where it is intensely deformed, fine cleavage has obliterated bedding fissility, and the noncalcareous shale resembles slate. This is called thin-cleaving shale, because thin-section study revealed that even the most closely spaced cleavage planes are jointlike fractures 0.02 to 0.06 mm apart and that there was no general orientation of argillaceous flakes parallel to cleavage.

Argillaceous limestone, fine-textured calcitite with a high proportion of argillaceous material, is characterized by varvelike lighter and darker laminae from less than 1 to 30 mm thick. Darker bands have a higher proportion of carbonaceous and argillaceous material. The proportion of detrital quartz silt varies from bed to bed and in places reaches a concentration sufficient to warrant calling the rock silty argillaceous limestone. This rock has platy to flaggy hard beds in the undeformed section in northwestern Grand Isle. Compression produced coarse fracture cleavage and eliminated all but the most prominent bedding separations.

Fine-grained limestone is calcilutite moderately rich in carbonaceous material, with occasional detrital grains of quartz silt or calcite silt (shell fragments).

Dolomitic siltstone is an argillaceous (about 25 per cent), quartz-silty (about 25 per cent) rock with about 50 per cent authigenic dolomite grains of silt size. Varying proportions of argil-

[1] The most widely used definition of the Hudson River group designated it as the age of the shales and sandstones of the Whetstone Gulf and Pulaski formations, of Cincinnatian age, overlying the Utica (Ruedemann, 1925, p. 8–10).

laceous and carbonaceous material form bedding laminae. Where dolomite reaches about 85 per cent, and quartz is rare to absent, the rock is fine-grained dolomite.

Table 1.—Autochthonous Trentonian Formations

Correlation	Formation	Lithology	Thickness
			(feet)
Upper Trenton ? (Collingwood-Gloucester)	Hathaway formation (new)	Argillite and bedded radiolarian chert, intensely deformed by submarine slides, with included boulders of chert, limestone, dolomite, sandstone, and graywacke	100+
Upper Trenton (Cobourg-Collingwood-Gloucester)	Iberville formation	Noncalcareous shale, rhythmically interbedded with thin quartz-silty dolomite beds, uniform, fine-grained dolomite beds, and in the lower part, with calcareous shale	1000–2000
	Stony Point formation	Predominantly calcareous shale in lower part, grading upward to predominantly quartz-silty argillaceous limestone	1000–1500
Sherman Fall	Cumberland Head formation	Interbedded calcareous shale and fine-grained homogeneous limestone	140+
	Shoreham member of Glens Falls formation	Thick-bedded homogeneous limestone, with shale partings	30+
Hull	Larrabee member of Glens Falls formation	Thin-bedded, somewhat shaly limestone	72
Rockland	Isle la Motte limestone	Heavy-ledged limestone	20+

Argillite, a highly indurated argillaceous rock, shows no bedding fissility.

Graywacke designates terrigenous clastic rock composed predominantly of sand-sized detrital grains of quartz, feldspar, weathered ferromagnesian silicates, slate, quartz aggregate, and chlorite, set in fine argillaceous and chloritic matrix.

Grain refers to any discrete mineral mass regardless of size, shape, or origin. No systematic study of microscopic grain size was made. Where nearly equidimensional grains were measured, the largest diameter was noted; for elongate grains, the longest and shortest dimensions.

In rock descriptions, color names and symbols are based on the National Research Council Rock Color Chart (1948).

Stony Point Formation

General.—Stony Point shale was designated by Ruedemann (1921a, p. 112–115) as "hard, splintery dark bluish-gray calcareous shale" near the Canadian border at Stony Point, 1½ miles south of Rouses Point, New York, on the west shore of Lake Champlain. Ruedemann correlated this shale faunally with upper Canajoharie shale of the Mohawk Valley in central New York. The Stony Point rests on interbedded calcareous shale and limestone first recognized by Perkins (1902, p. 167) as transitional between "Trenton" (Shoreham) limestone and overlying shales. Transition beds were named Cumberland Head shale by Cushing (1905, p. 376, Pl. 13). Ruedemann (1921a, p. 114) and Kay (1937, p. 274–275) further clarified the relation of Stony Point to the Cumberland Head along the west shore of Grand Isle south of Gordon Landing.

Base of Stony Point.—Deposition was apparently continuous from Cumberland Head upward into Stony Point. Choice of contact is somewhat arbitrary and is based on lithologic criteria in shore outcrops 0.55 mile south of the breakwater at Gordon Landing, and 0.6 mile north of the north edge of Wilcox Bay. South of Gordon Landing about 140 feet of Cumberland Head beds dips gently northward between a strike-slip fault and the base of the Stony

Point. Cumberland Head is composed of 3- to 12-inch beds of light-olive-gray (5Y6/1) weathering, dark-gray (N3), fine-grained limestone usually with undulating bedding surfaces, interbedded with half-an-inch to 12-inch layers

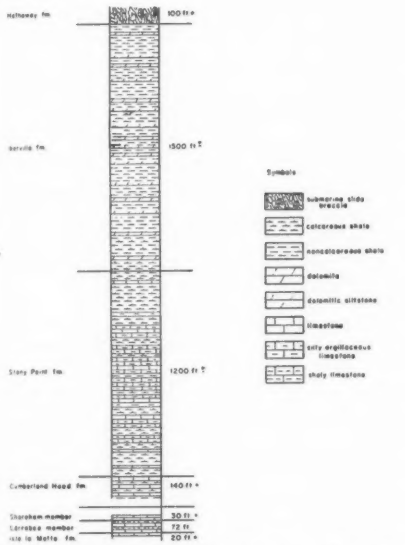

FIGURE 1.—COLUMNAR SECTION OF TRENTON FORMATIONS

Central lowland of the northern Champlain Valley.

of black calcareous shale. Some limestone beds as thick as 8 inches grade laterally into shale within 50 feet. Less than one-third of the Cumberland Head is composed of more than 50 per cent shale, and about half has more than 60 per cent limestone beds. Some units as much as 15 feet thick are composed of 80 per cent limestone beds. The proportion of shale increases gradually but not uniformly upward.

Lower Stony Point south of Gordon Landing has a greater proportion of black calcareous shale and does not have the wavy, irregular bedding found in the Cumberland Head. Dark-gray limestone beds (1 to 12 inches) are common and in two places make up more than 80 per cent of rock units 9 feet thick. The upper part of this section has units of black calcareous shale as thick as 42 feet, with occasional (every 6 to 8 feet) thin, dark-gray limestone beds (a quarter of an inch to 1 inch). Lenticular, radial concretions of pyrite, averaging half an inch by 1 inch, are fairly common in a thin zone of fine, disseminated pyrite along the middle of some limestone interbeds. The Stony Point section south of Gordon Landing, 215 feet thick (Kay, 1937, p. 274), is cut off at the north (upper) end by a fault.

Northwest Grand Isle section.—The thickest and least deformed section of Stony Point measured in detail begins 0.6 mile north of Wilcox Bay and extends for 1.8 miles northward along the west shore-line bluffs of Grand Isle. The beds dip gently northward. This section totaled 634.6 feet, including eight covered zones equivalent to 111.3 feet. No structure in rock bordering the covered areas indicates faulting. It is impossible to tell whether two finely brecciated and slickensided bedding-shear zones 532.5 feet (5 inches thick) and 564.3 feet (6 inches thick) above the base of the Stony Point cause any stratigraphic displacement; but the gentle uniform dip and absence of disturbance above and below the zones suggests only minor effect on the thickness of the section. This section is described in detail in Appendix, Section 1.

A few gross vertical lithologic variations recognizable throughout the field area are found in the northwest Grand Isle section. The lower 200 feet is intercalated dark-gray (N3) calcareous shale with light-olive-gray (5Y6/1) weathering, dark-gray (N3) fine-grained limestone in occasional thin beds, about 71 per cent shale, and closely resembles the section south of Gordon Landing. From 235 feet above the base, the percentage of calcareous shale decreases upward. Olive-gray (5Y4/1) to light-olive-gray (5Y6/1) weathering, dark-gray (N3) argillaceous limestone appears in increasing proportion through the remaining 400 feet of the section, where the estimated percentages of the lithologies are: argillaceous limestone, commonly silty, 66 per cent; calcareous shale, 29 per cent; fine-grained limestone beds, 5 per cent.

Argillaceous limestone commonly occurs in thin, even beds (a quarter of an inch to three-quarters of an inch) with fine lighter- and darker-gray laminae, but occasional beds reach 10 inches. Thicker-bedded zones suggest cyclic deposition: from calcareous shale (1 to 4 inches) upward through 4 to 5 inches of laminated

argillaceous limestone to a 1- to 3-inch bed of fine-grained limestone; then through 4 to 5 inches of argillaceous limestone to 1 to 4 inches of calcareous shale. Where the interval between the calcareous shale beds is thinner, the fine-grained limestone bed in the middle is absent. The proportion of silt and argillaceous material in harder argillaceous limestone varies greatly. Intricate, fine cross-bedding occurs in four thin zones 434, 458, 532, and 620 feet above the base of Stony Point. The lowest and uppermost of these zones have ripple-drift bedding (Bucher, 1919, p. 155; Sorby, 1908, p. 184) indicating currents flowing N. 20° E. and N. 40° E., respectively. Ripples in the lowest zone have a wave length of 18 inches and an amplitude of 1½ inches; and in the upper zone, a wave length of 12 inches and an amplitude of 1 inch. These amplitudes have probably been decreased by compaction of fine-grained sediment.

Light-olive-gray (5Y6/1) weathering, dark-gray (N3), fine-grained limestone occurs throughout the section at intervals of 3 inches to 20 feet in beds half an inch to 1 foot thick, which are occasionally laminated.

A distinctive bed of medium-light-gray (N6) weathering, medium-gray (N5), fine-grained crystalline limestone, ranging in thickness from half an inch to 1½ inches, and occasionally lensing out, occurs 532 feet above the base of the Stony Point in the northwest Grand Isle section. Four such beds were noted in the 258-foot section of light- and dark-gray-banded, laminated argillaceous limestone on the promontory west of Coon Point in south Alburg. Throughout the field area they are associated with the argillaceous limestone which appears in the upper part of the northwest Grand Isle section.

Small pyrite concretions, common in the upper and lower parts of the section, are distributed along occasional thin bedding zones.

A 29-foot section of interbedded calcareous shale (40 per cent), silty argillaceous limestone (55 per cent), and fine-grained limestone in 1- to 4-inch beds (5 per cent), exposed three-quarters of a mile south of Long Point, northwestern Grand Isle, is separated from the north end of the measured section by 0.7 mile of cover. This outcrop is about 240 feet stratigraphically above the top of the 634.6-foot section, unless there is some disturbance between them. Whether two ripple-marked surfaces on beds of silty argillaceous limestone show wave or current ripples could not be determined because compaction has flattened them. Their wave lengths are about 3½ inches, and they trend N. 46° W. and N. 80° W. In contrast with the 634.6-foot section, this outcrop is fossiliferous, with *Triarthrus*-like fragments, *Diplograptus amplexicaulus* (Hall), and *Lasiograptus eucharis* (Hall). The faunule resembles Ruedemann's in the nearly 75 feet of silty argillaceous limestone at Stony Point (1921a, p. 112) and in west-central Vermont at Arnolds Bay in Panton (1921a, p. 110), which Ruedemann correlated with the upper Canajoharie.

South Alburg section.—On an unnamed promontory west of Coon Point in southern Alburg, 258 feet of laminated argillaceous limestone was measured (Appendix, Section 2) and appears to be, at least in part, younger than the northwest Grand Isle section. There are only three lithologic types. Olive-gray (5Y4/1) weathering, dark-gray (N3) argillaceous limestone, frequently silty, with light-olive-gray (5Y6/1) weathering bands and laminae, composes 96 per cent of the section, occurring in units of 3 inches to 40 feet. Light-olive-gray weathering (5Y6/1), dark-gray (N3), fine-grained limestone, composing 4 per cent of the section, occurs in beds of half an inch to 12 inches, separated by 3-inch to 40-foot units of silty argillaceous limestone. Four beds of medium light-gray (N6) weathering, medium-gray, fine-grained crystalline limestone a quarter of an inch to 1¼ inches thick occur 44.9, 82.3, 170.5, and 218.6 feet above the base of the measured section. Pyrite concretions are common. This section correlates well lithologically with the upper portion of the northwest Grand Isle section and with the section at Stony Point. It is less fossiliferous than Ruedemann's Stony Point locality and has only a profusion of *Triarthrus becki* (Green) scattered throughout with a few unidentifiable graptolite fragments.

Accessible exposures roughly correlative with the south Alburg measured section bearing *Triarthrus*, are: (1) the road cut on U. S. Highway 2, 0.3 mile south of the Carrying Place, North Hero, where varvelike lamination in argillaceous limestone is well shown; (2) South

Hero on a hillside south of U. S. Highway 2, 1.8 miles S. 5° W. from Kibbie Point; (3) a quarry on the west side of U. S. Highway 2 in Alburg, 3.3 miles north of the west end of the Alburg-North Hero bridge.

A fine-grained crystalline limestone bed is exposed in a bluff of calcareous shale and argillaceous limestone on the west shore of southern Alburg, 2.15 miles north of Coon Point. This bed is unusual for its thickness, about 5 inches, for current ripples on its upper surface, and for the overlying 6-inch bed of noncalcareous shale. The current ripples have a wave length of 5–7 inches, an amplitude of about 1 inch, strike N. 68° W. and indicate a northeastward-flowing current, corresponding with ripple drift in northwest Grand Isle. The overlying 6-inch bed of moderate-brown (5YR3/4) to dusky-moderate-brown (5YR2/5) weathering, dark-gray (N3) noncalcareous shale is brittle and hard and resembles nothing else in the section.

Upper portion of Stony Point.—On the east shore of North Hero, north from the Carrying Place, is 2 miles of almost continuous shore-line bluff of calcareous shale and laminated argillaceous limestone with thin beds of fine-grained limestone and two thin fine-grained crystalline limestone beds similar to those near Coon Point and those in the upper part of northwest Grand Isle section. Two gently northeastward-plunging folds indicate that younger beds are northward. These younger beds show an increasing proportion of calcareous shale. On the shore at Elm Point Farm, 1.4 miles S. 37° W. from Long Point, North Hero, interbedded argillaceous limestone and calcareous shale, predominantly limestone, are overlain by 239 feet of calcareous shale presumably younger than the youngest beds of the northwest Grand Isle section.

Beds transitional from the Stony Point to the Iberville formations are found still farther northeastward, 0.55 mile S. 85° W. of Long Point, North Hero, where grayish-black shale with weathered surfaces show light- and dark-gray laminae caused by narrow bands of thin leached-lime coating. Thus, the shale is composed of calcareous and noncalcareous laminae. Brownish-gray (5YR4/1) weathering silty layers, about a quarter of an inch thick, lie 2 to 4 inches apart. There is one 4-inch, moderate-yellowish-brown (10YR5/4) weathering, medium- to dark-gray (N5–N3), fine-grained, laminated dolomite bed in this outcrop. Noncalcareous shale, dolomitic siltstone, and dolomite beds, even where interbedded with calcareous shale, characterize the Iberville formation. The base of the Iberville has been placed at the bottom of the transition zone, where noncalcareous shale and thin dolomitic beds appear in the section. The actual contact is not exposed in North Hero. No reliable estimate can be made of the thickness of the covered section between these older Iberville beds and the 239-foot shale section at Elm Point Farm.

Thickness of Stony Point.—Total measured thickness of the Stony Point formation is 873.6 feet. This is the sum of the northwest Grand Isle section and the shale at Elm Point Farm; the shale is almost certainly younger than the youngest beds in Grand Isle. This is a minimum thickness because the measured section in South Alburg probably includes beds younger than the Grand Isle section. In the Senigon well, near Noyan, Quebec, about 4 miles north of the international boundary at Alburg, shale apparently equivalent to the Stony Point is 924 feet thick (Clark and Strachan, 1955, p. 687–689).

Thin sections of Stony Point.—Nine thin sections of the Stony Point were examined, four from beds in the measured section on the promontory west of Coon Point, south Alburg, and five from fine-grained crystalline limestone beds elsewhere.

SILTY ARGILLACEOUS LIMESTONE: Silty argillaceous limestone, from the lower part of the south Alburg measured section, is dark-gray (N3), with olive-gray (5Y4/1) and light-olive-gray (5Y6/1) weathering strata and laminae from less than 1 to 30 mm. Between the thicker strata commonly is a transition zone of about 3 mm with alternating thin laminae (0.6 mm ±). The darker rock has dark-brown, opaque to translucent carbonaceous material densely scattered as irregular fine clots and streamers parallel to bedding. The lighter zones have much less of this. About 75 per cent of the rock is fine-grained calcite (0.003 mm and smaller), with occasional euhedral and irregular grains as large as 0.032 mm. Quartz is occasionally found in both the lighter and darker bands, commonly as small (0.004 to 0.025 mm.) irregular to angular grains, averaging about 0.011 mm. Some anhedra have minute carbonate inclusions and seem authigenic. Only small angular chips appear detrital. Pyrite is common, disseminated as fine

grains (0.01 mm), with occasional larger irregular to globular grains (0.032 mm). Certain zones richer in fossil fragments as large as 0.29 by 0.016 mm occur in both darker and lighter strata, but the proportion of rock made up of such fragments is small.

FINE-GRAINED LIMESTONE: The dark-gray (N3), weathering light-olive-gray (5Y6/1), limestone from the lower part of the south Alburg section is predominantly a groundmass of extremely fine-grained (0.001 to 0.003 mm) calcite. Disseminated through the groundmass is finely divided, dark-brown, translucent to opaque carbonaceous material concentrated in streamers parallel to bedding. Scattered irregular and angular quartz grains average about 0.011 mm. Pyrite is widely disseminated as small anhedra (0.004 mm, occasionally as large as 0.065 mm). Pyritic replaces some of the small (0.07 mm) organic fragments. Two small fossil fragments in the thin section were replaced in geode fashion by a rim of calcite crystals growing inward, while the interior was filled with quartz.

FINE-GRAINED CRYSTALLINE LIMESTONE: The medium-gray (N5), weathering medium-light-gray (N6), limestone was selected for thin sections from beds 170.5 and 218.6 feet above the base of the south Alburg section; 2.15 miles north of Coon Point, on the west shore of south Alburg; 0.75 mile north of Hazen Point, on the west shore of North Hero; 1.5 miles north of the Carrying Place on the east shore of North Hero; 0.2 mile west of Ladd Point, northern Grand Isle; 0.35 mile south of Beans Point on the east shore of Lake Champlain. All specimens are similar except that from the bed near Beans Point, which is near the Champlain thrust and is mylonitized.

The fine-grained crystalline limestone is about 70 per cent calcite, in interlocking, sutured grains which in most specimens are 0.16–0.24 mm across; in some beds calcite-grain diameters are 0.11 to 0.13 mm. By ordinary light the sections show peculiar fragmental structure with curved spicules and larger angular grains with sharply arcuate edges, suggesting palimpsest bogen structure. Under crossed nicols the calcite grain boundaries rarely have these forms, though one (0.273 mm long) strongly resembled a pyroclastic shard.

The crystalline limestone is rather richly scattered with small (0.015–0.03 mm long) euhedral feldspar crystals, occasionally as large as 0.018 by 0.05 mm. Inclusions of calcite and scattered unidentifiable specks are common in the feldspar. Although symmetrical extinction on twin lamellae is very rare, one good example measured 13°. These twinned crystals are probably albite. Many feldspar crystals, particularly the smaller ones, are untwinned, and their habit is characteristic of adularia. In view of predominantly euhedral shape, evidence of replacement of calcite by feldspar (inclusions and occasional serrate borders), and absence of detrital cores, it seems that all or almost all the feldspar is authigenic. The origin of similar occurrences of authigenic feldspar in sediments is discussed by Tester and Atwater (1934, p. 23–31), Goldich (1934, p. 89–95), and Honess and Jeffries (1940, p. 12–18).

Rare grains with low relief, index greater than balsam, and weak birefringence suggest quartz. Detrital quartz, common in associated beds, is absent. Detrital zircon and rutile are rare. Pyrite is common, disseminated in fine irregular grains (0.016 mm ±) and occasionally as cubes up to 0.07 mm. Brownish chalcedony is fairly common in grains up to 0.07 mm, and contains abundant calcite inclusions which suggests that it has replaced calcite. Chalcedony sometimes occurs in small, irregular patches between calcite grains.

The crystalline limestone mylonite bed near Beans Point is macroscopically identical to the other beds. In thin section, however, calcite grains as large as 0.11 by 0.073 mm have curved secondary twin lamellae, are cut by curved shear planes, and are imbedded in finely crushed calcite. Interlocking grain pattern with sutured boundaries found in other such beds is missing in this rock. Fibrous, elongate, bent grains do not extinguish under crossed nicols. Feldspar grains, somewhat less common than in the other beds, are rarely euhedral, and often fractured. Quartz, though rare, is more common than in other beds. Some grains are as large as 0.073 mm, commonly well rounded, with strain shadows. Pyrite is common, finely disseminated, with occasional lenticular masses as large as 0.032 by 0.114 mm. Brown, translucent to opaque carbonaceous material, clotted and streamed through the thin section, is occasionally associated with finely divided pyrite.

Soluble proportion of Stony Point.—Specimens of Stony Point lithologic types were crushed and tested with dilute hydrochloric acid for proportion of soluble constituents.

	Number of Specimens	Range of Solubility	Av. Solubility
		(Per cent)	*(Per cent)*
Silty argillaceous limestone	4	68–77	72
Calcareous shale	1	...	48
Fine-grained limestone	3	73–82	78
Fine-grained crystalline limestone	2	68–72	70

Iberville Formation

General.—Detailed description of the Iberville formation, named by Clark (1934, p. 5), is postponed until publication of Geology of the Lacolle quadrangle by T. H. Clark and H. W. McGerrigle. Professor Clark kindly allowed the writer to examine his field maps and the manuscript from which the following pertinent data are summarized. The Iberville is named for its wide outcrop belt in Iberville County, southern Quebec. In the Lacolle quadrangle, directly north of the Canadian border in central Champlain Valley, Clark and McGerrigle distinguished the same two lithologic sequences present in Vermont: the older is Stony Point; and the younger, Iberville, which includes "dark-gray, thin-cleaving argillaceous shales, much sheared, with occasional siliceous and calcareous beds interstratified". Clark and McGerrigle found that the Iberville contains no fossils preserved well enough to be of value, but in consideration of Stony Point faunas examined by Ruedemann (1921b, p. 96–99), and of Stony Point collections made in southern Quebec, they conclude that the Iberville must be Utica in age. Clark found (1939, p. 582) that the Iberville is 1000 to 2000 feet thick in its type area, and it has to grade northward through the Island of Montreal into thinner (400 feet) typical Utica shales. In the 2296-foot Senigon well, Clark and Strachan (1955, p. 687–688) found about 1200 feet of Utica shale, presumably Iberville.

In the area studied by the present author the Iberville is usually more intensely deformed than the Stony Point or any other underlying formation principally because most outcrop areas are nearer major thrusts. Its exposures are more limited because of its greater apparent susceptibility to weathering and erosion. Broad areas where Iberville should be exposed are covered with marsh, overburden, or northeastern Lake Champlain.

Lower Iberville contains intercalated calcareous and noncalcareous shale and occasional beds of dolomite, showing a gradation from underlying Stony Point. Sections of lower Iberville are: (1) the shore southeast of Kibbee Point, South Hero; (2) Appletree Point, 3 miles northwest of Burlington; (3) Savage Island, east of Grand Isle.

Upper Iberville is composed of interbeds of noncalcareous shale (1–12 inches) and laminated dolomitic siltstone (a quarter of an inch to 10 inches) in cyclic repetition through hundreds of feet, with occasional beds (1½–12 inches) of fine-grained dolomite. Exposures of upper Iberville are: (1) two points north of Cary Bay, North Hero; (2) in a quarry used for road metal and rip-rap 1.6 miles S. 10° E. from the east end of the North Hero-Alburg bridge; (3) in a small quarry and along the shore, about 1.2 miles north of Long Point, North Hero; (4) Clark Point on Hog Island in west Swanton; (5) west shore of Woods Island, west St. Albans. Younger age of these beds is inferred from their structural position in North Hero, where they lie in a deep syncline north of Cary Bay, and on a series of folds down-plunge (northeastward) from the outcrop of Stony Point-Iberville transition beds west of Long Point.

Lower Iberville southeast from Kibbee Point, South Hero.—At Kibbee Point, interbedded shale and argillaceous limestone of Stony Point formation are tightly compressed and sheared on an anticlinal crest, and form high bluffs. Southeastward toward a syncline a considerable thickness of calcareous shale corresponds to that at Elm Point Farm, North Hero. Transition beds at the base of the Iberville, a quarter of a mile southeast of Kibbee Point, are about 80 per cent dark-gray (N3) calcareous shale with thin beds (half an inch) of dark-gray noncalcareous shale. They have thin cleavage, and weathered cleavage surfaces show a banding of leached lime similar to the outcrop 0.55 mile S. 85° W. of Long Point, North Hero.

Beds of typical Iberville crop out about 100 feet farther southeast and dip 20°–40° SE. Estimated lithologic proportions in this part of the section are: noncalcareous shale, 60 per cent; calcareous shale, 35 per cent; occasional moderate-yellowish-brown (10YR5/4) weathering, dark-gray, fine-grained, laminated or homogeneous dolomite, in beds as thick as 8 inches, 5 per cent. About 150 feet southeastward, still higher in the stratigraphic section, calcareous shale almost disappears, and individual noncalcareous thin-cleaving shale beds attain 18 inches and average about 8 inches.

From here to the fault half a mile southeast of Kibbee Point shale is almost entirely noncal-

careous, with occasional argillaceous dolomite beds having fracture cleavage (2 to 12 inches), hard fine-grained dolomite beds (2 to 8 inches), and laminated dolomitic siltstone beds (1 to 8 inches) some of which have ripple-drift cross-lamination. Higher in this section are a few dolomitic quartz sandstone beds as thick as 8 inches ("curious dike-like bands of highly siliceous rocks, weathering rust-brown," of Perkins, 1902, p. 132).

Along this part of the shore the southeastward dip steepens to vertical, and the folding and occasional shearing, as well as frequent cover, made it impracticable to measure the section. It is estimated to be more than 300 feet thick.

Lower Iberville, Appletree Point, Burlington.—A 71.6-foot section of transition beds, presumably at the base of the Iberville, is exposed on the east side of Appletree Point, 3 miles northwest of Burlington, Vermont. Because of isolation of this outcrop and complexity of structure, its stratigraphic position cannot be located accurately. It is composed of interbedded: dark-gray (N3) noncalcareous shale with very thin cleavage, in beds of half an inch to 8 inches; grayish-black (N2) calcareous shale with thin cleavage, in beds of half an inch to 10 inches; light-olive-gray (5Y6/1) weathering, dark-gray (N3), fine-grained limestone, occasionally laminated or argillaceous or both, in beds a quarter of an inch to 5 inches; moderate-yellowish-brown (10YR5/4) weathering, dark-gray (N3), fine-grained dolomite, occasionally laminated and cross-bedded, in beds of half an inch to 10 inches. Approximate proportions are: calcareous shale, 32 per cent; noncalcareous shale, 40 per cent; limestone, 20 per cent; dolomite, 8 per cent. Northward along shoreline bluffs from Appletree Point no section measurements were possible because of intense folding and minor thrusting, but the lithology is very similar, with varying proportions of calcareous and noncalcareous shale.

Lower Iberville, Savage Island.—Higher stratigraphic position of Iberville relative to Stony Point is confirmed on Savage Island. Rocks of the northeastern part of the island are estimated to be 60 per cent gray noncalcareous shale with very thin cleavage, in beds of 1 to 12 inches; 25 per cent gritty shale with thin cleavage in similar beds, which effervesce faintly with dilute hydrochloric acid; and 15 per cent moderate-yellowish-brown (10YR5/4) weathering, fine-grained dolomite beds half an inch to an inch thick, occurring every 6 to 24 inches. These are typically lower Iberville. A small slate quarry here was worked briefly about 75 years ago, according to local mainland residents. This section lies along the axis of an overturned (NW), thrust-faulted syncline plunging northeastward. The western, southern, and southeastern portions of the island have outcrops of typical Stony Point argillaceous limestone and calcareous shale.

Upper Iberville, Clark Point, west Swanton.—At Clark Point a 303.7-foot section of younger Iberville was measured and is described in detail in the Appendix, Section 3. It is 97.8 per cent thin cleaving noncalcareous shale with thin laminated dolomitic siltstone interbeds (about 4 per cent of these beds are siltstone) and 2.2 per cent laminated or homogeneous fine-grained dolomite beds up to 14 inches thick, occurring at intervals of 3 to 50 feet. The shale and siltstone are shown in Plate 2.

Upper Iberville, Woods Island, west St. Albans.—On the west shore of Woods Island two sections of similar Iberville beds totaled 369.1 and 363.2 feet (Appendix, Section 4). Both sections have the same lithologic proportions: 98.2 per cent thin-cleaving shale with laminated dolomitic siltstone interbeds (about 4 per cent) and 1.8 per cent laminated or homogeneous fine-grained dolomite in beds 1½ to 12 inches thick, occurring 1 to 50 feet apart. The top of the 369.1-foot section is separated from the bottom of the 363.2-foot section by 1000 feet of shoreline cover which parallels the strike of the beds. If beds are not displaced beneath the cover, there should be approximate correlation between the top of the older section and the bottom of the younger. Almost identical strikes and dips at both ends of the cover support such an assumption.

Thickness of Iberville.—If the inferred sequence on Woods Island is correct, the total of the two sections, 732.3 feet, is a minimum limit for the Iberville. This does not include the transition beds at the base of the Iberville, which unfortunately could not be measured. Including the transition beds, a thickness of

1000 feet is certainly credible, and it might be greater. This sustains Clark's estimate of 1000 to 2000 feet in southern Quebec (1939, p. 582) and corresponds well with the 1200 feet of Utica shale in Senigon Well 30 miles southeast of Montreal (Clark and Strachan, 1955, p. 687–688).

Primary structures in Iberville.—Iberville shale and dolomitic siltstone show remarkable rhythmic bedding. The base of each cycle is a slightly scoured surface (Pl. 2, fig. 1) on which a thin bed (0.25–0.75 inch) of yellowish-brown (10YR5/4) weathering, medium dark-gray (N4), laminated dolomitic siltstone was deposited. Siltstone layers become finer-grained upward with decreasing quartz and dolomite, and increasing argillaceous and carbonaceous material, and grade into dark-gray (N3) noncalcareous thin-cleaving shale beds 1–4 inches thick. Usually at the top is an eighth to three-quarters of an inch of grayish-black (N2) shale interlaminated with the dark gray. Occasionally the dolomitic siltstone may be missing at the bottom of the cycle, or the grayish-black shale laminae missing at the top.

Moderate-yellowish-brown (10YR5/4) weathering, medium-dark-gray (N4), fine-grained dolomite beds (2–4 inches) occur at intervals of 3 to 50 feet. These frequently have dark-yellowish-brown (10YR4/2) weathering, dark-gray (N3), laminated zones an inch or more thick, which occasionally constitute the entire bed. The lower half of such a bed is occasionally laminated dolomitic siltstone (laminae 0.1–1.0 mm), commonly with current cross-lamination, while the upper portion is dense, homogeneous fine-grained dolomite.

There are a few gradations upward and laterally from dense fine-grained dolomite to thin-cleaving shale. An 8-inch bed of fine-grained laminated dolomite 86.6 feet below the top of the Clark Point section grades upward into shale, and laterally as well, within 50 feet along the strike. On the south shore of St. Albans Point a similar dolomite bed (1 foot) grades into shale laterally within 4 feet (Pl. 3, fig. 2). Thin-section study sheds little light on the origin of this lateral gradation.

Ripple-drift cross-lamination (Bucher, 1919, p. 155; Sorby, 1908, p. 184), a common feature of dolomitic siltstone beds of the Iberville, is particularly well shown on the easternmost point of Burton Island where 15 moderate-yellowish-brown (10YR5/4) weathering, dark-gray (N3) dolomitic siltstone beds with ripple drift, 2–4 inches thick, occur in a 32.9-foot section. In some beds only a single set of ripples was built. In others down-current ripple drift continued long enough for one ripple to climb the stoss slope of the next, to produce commonly two, and occasionally three or four tiered beds. The number of tiers is not constant even along the same bed. In some ripples deposition was so rapid that laminae were laid down on both stoss and lee slopes (Pl. 2, fig. 3). Other ripples show scour on the stoss slope with deposition on the lee. Coarser material was laid down on the stoss slope when deposition took place there, and at the top of the lee slope. Lower on the lee slope there is gradation to very fine sediment. Thus, it often appears that ripples drift upward and fade into the overlying shale beds (Pl. 2, fig. 3). In photographs the lighter tone reflects dolomitization, which appears only in the silty laminae.

With few exceptions the ripples were exposed on cleavage surfaces where it was impossible to determine crest-line bearings. Dips of the laminae are apparent dips with no possibility of correction. Since wave-length measurements are similarly affected, and amplitude measurements are not, vertical form indices (Bucher, 1919, p. 154) computed from them would be misleading. Amplitudes varied from 1 to 3 inches in different beds on Burton Island. Iberville beds in the syncline southeast of Kibbie Point, South Hero, have even larger ripples in one bed, with amplitudes of 5 inches. Length measurements of the stoss and lee sides are affected to the same degree, and valid horizontal form indices (Bucher, 1919, p. 154) were obtained from measurements made on cleavage surfaces. Nine ripples measured on northeastern Burton Island yielded horizontal form indices ranging from 2 to 5.3. Four ripples in sequence on a single bed had horizontal form indices of 3, 3.5, 5.3, and 4.

It was possible to determine true wave lengths on only one set of ripples, on the shore 0.3 mile northeast of the southern tip of Burton Island. Five ripples in sequence on the same bed measured 30, 36, 46, 45, and 48 inches.

These are extraordinary wave lengths for ripples in such fine-grained sediment. Bucher (1919, p. 264) states that "large current-ripples are found only on rocks of relatively coarse grain, as conglomerate sands . . . or fragmental limestones, never on fine grain sediments." His summary (p. 156–159) of Hahmann's evaluation of factors influencing wave length does not throw any light on this problem. Hahmann found:

(1) Wave length increases with current velocity. As there is no evidence of wave action or wave-induced turbulence in the Iberville, and its ripples are so large, one must assume fast currents in fairly deep water. Fast currents normally would not deposit sediment so fine.

(2) Wave length increases with increasing grain size. This merely emphasizes the paradox of long ripples in fine-grained sediments.

(3) Wave length is inversely proportional to viscosity of water as influenced by salinity and temperature. If viscosity increased by turbidity has the same result, this should decrease rather than lengthen the wave lengths of ripples formed.

(4) Within the narrow limits of his experimental work, Hahmann found no influence of water depth upon wave length. There must be some condition of sedimentary environment under which these beds were deposited which was not considered in Hahmann's analysis. This will be discussed below with reference to the graded bedding of the Iberville.

Direction of crest line of ripples could be measured on only four beds. Compensated for bedding dip where necessary, the inferred current directions noted are:

Eastern point of Burton Island	S. 23° W.
0.3 mile northeast of south tip, Burton Island	S. 45° W.
Point north of center of Cary Bay, North Hero	S. 33° W.
Quarry 1.6 mile S. 10° E. of North Hero-Alburg bridge	S. 15° E.

The general current direction could be determined, however, in 24 beds at six localities throughout the field area of the Iberville. Only currents flowing toward south to west are indicated. This contrasts with the two Stony Point beds in the northeast Grand Isle section, indicating currents flowing N. 20° E. and N. 40° E., and the Stony Point fine-grained crystalline limestone bed on the southwest shore of Alburg, which indicated a current flowing N. 22° E. Current data from the Stony Point are few, and data from the Iberville are not so complete as one might wish, but they are consistent. There is a suggestion of a major change in source of sediments from the Stony Point to the Iberville which can be based upon current direction as well as lithology. A study of heavy-mineral concentrations in the two formations might be very revealing. Although current ripple trends in older beds are not noted for eastern Pennsylvania, the Martinsburg formation cross-bedding everywhere indicates westward-flowing currents (Behre, 1927, p. 84). This formation is roughly correlative with the Iberville.

Evidence for deposition by turbidity currents is found in the Iberville. Six dark-greenish-gray (5GY4/1) weathering, medium-dark-gray (N4) graded siltstone beds 5–10 inches thick occur at 2- to 9-foot intervals in the 32.9-foot section measured on northeastern Burton Island. Lower portions of these beds are rather coarse, with many flakes of shale 1 mm long. Contact with underlying shale beds shows no evidence of scour. They grade finer upward but in places are laminated above the lower third. One bed has several large angular shale fragments in its mid-portion; the largest is 5 by 1¼ inches (Pl. 2, fig. 3). The associated rumpled structure characteristic of the mid-portions of these beds is strikingly similar to that produced experimentally by Kuenen from turbidity currents (Kuenen and Migliorini, 1950, Pls. 3, 4). Upward, lamination is more marked, and the beds are topped with drift ripples. These strata suggest: (1) rapid deposition which slowed as turbidity-current density decreased because of mixing with sea water; and (2) that turbidity-current deposition changed to normal current deposition on drift ripples. Kuenen recognized this possibility (1950, p. 98) and has discussed in some detail the relationship of current bedding to graded bedding (1953, p. 1051).

Henderson (1936, p. 497) points out that, although Bailey (1930, p. 85) considered graded bedding and current bedding the distinguishing marks of two different facies, the Ardwell beds of the Girvan District, Scotland, "not infre-

quently show an association of the two particular types of bedding within the limits of a single bed. In such a case the current bedding is on a small scale relative to the thickness of the bed as a whole, and it interrupts and complicates the grading." This description applies to some graded beds of the Iberville. In one bed (Pl. 3, fig. 1), the lower 2 inches is graded siltstone with flecks of shale, complicated by crude lamination. The central portion (2 inches ±) is darker, finer-grained silty mudstone, capped by laminated dolomitic siltstone with ripple drift. Subaqueous gliding has caused slipping in the lower part of the bed, and flowage of the top layer into the underlying mud has resulted in flow-cast (Shrock, 1948, p. 157) or load-case (Kuenen, 1953, p. 1658) structures. In many features the Iberville is strikingly similar to the Ardwell beds Henderson described. Both are interbedded coarse clastic and mudstone strata, with grading in coarser beds commonly complicated by current bedding, though in the Iberville the sands seem to be predominantly finer-grained, and in smaller proportion. Both have cyclic deposition patterns of similar nature. In both, the thickness of each cycle ranges from about 1 to 18 inches, but mostly 1 to 5 inches, through a thickness of hundreds of feet. Throughout both formations intrastratal crumpling is found, and, if the Hathaway formation be included, both have breccia and coarse grit, with large-scale submarine slumping. In the Clark Point section of 303.7 feet, about 1215 cycles are estimated. In the Woods Island sections a similar estimate is about 2200 for 732.3 feet. This compares with 4000 graded beds in the 1000 feet of the Ardwell group which Henderson measured (1936, p. 497). Similar regular alternation of bedding is typical of the lowest member of the Martinsburg formation in Pennsylvania (Behre, 1927, p. 19), where there are also some minor folds probably formed while the muds were being deposited (p. 23).

Following Bailey (1930, p. 89), Henderson (1936, p. 497) suggests that bottom disturbances and tsunamis resulting from earthquakes account for each cycle. This explanation seems equally appropriate for the Iberville, with the added refinement of turbidity currents to account for transport and deposition (Kuenen and Migliorini, 1950, p. 122–126). In a tectonically active region earthquakes occur frequently enough to account for periodic mild disturbances resulting in turbidity currents. The Taconic orogeny during later Ordovician time and the Vermontian uplift (Kay, 1937, p. 290) at the close of Sherman Fall time suggest that this region was tectonically active during Iberville deposition. Whether earthquakes would recur with the great regularity indicated by the bedding is another question. Barrell (1917, p. 804) accounted for the similar beds in the Martinsburg with climatic rhythms of varying storm intensity, periodically scouring and agitating the bottom sediments by wave action. In the Iberville there are no wave ripple marks; on the contrary, deep water is suggested by the very fine grain size and lack of apparent disturbance after deposition. Rich (1950, p. 719; 1951, p. 6) points out that the Iberville type of bedding is characteristic of clino, or "slope" environments down which turbid currents flow from shallower water after turbulence during severe storms. Thus, the Iberville rhythms may represent a climatic cycle which induced periodic turbidity currents.

Iberville siltstone beds probably are peripheral deposits where turbulence has diluted the turbid current with overlying sea water. Commonly the beds become finer-grained upward through a series of thin laminae, rather than showing the grading typical of experimentally produced turbidity currents. Such peripheral deposits might be very broad, extending from the bases of slopes, aided in their distribution by slow, deep-water currents.

Thin sections of Iberville.—Thin sections were made from the thin-cleaving shales, the dolomitic siltstone, and the fine-grained dolomite beds of the Iberville formation.

THIN-CLEAVING SHALE: The shale is composed chiefly of finely divided tabular particles of argillaceous material of low relief and very weak birefringence. Scattered throughout the groundmass are: moderately abundant angular to subangular quartz grains (0.003 to 0.04 mm, averaging about 0.007 mm); moderately abundant sericite flakes (0.001 by 0.003 mm, to 0.004 by 0.04 mm); occasional tiny carbonate rhombs (0.003 mm) and irregular larger grains (0.025 mm); frequent pyrite anhedra about 0.004 mm across, occasionally as large as 0.08 mm; moderately abundant opaque black to brownish flakes (averaging 0.008 by 0.07

mm) of carbonaceous material. Minute particles in the groundmass appear to have random orientation, but more elongate quartz grains, sericite flakes, and the carbonaceous fragments commonly parallel bedding. Grayish-black (N2) zones have a greater concentration of small carbonate grains (0.01 mm ±) and much more carbonaceous material.

LAMINATED DOLOMITIC SILTSTONE: An estimated 50 per cent of this siltstone is carbonate grains (0.05 mm ±), presumably iron-rich dolomite because of the high relief and indices of refraction both of which are greater than the refractive index of Canada balsam. Hand specimens effervesce weakly with dilute hydrochloric acid. Much of the dolomite is euhedral. Some crystals show zoned impurities. An estimated 25 per cent of the rock is angular to rounded quartz grains (0.038 mm ±). The remaining 25 per cent is a very fine-grained groundmass of argillaceous material with irregular masses of brownish isotropic to opaque carbonaceous material. This gradually increases in proportion upward in each lamina, as well as in each laminated bed. Accessory minerals in very small grains (0.02 mm ±) are flakes of biotite and sericite, plagioclase, leucoxene, and occasional small rounded grains of zircon, rutile, and spinel. Pyrite (0.004 to 0.11 mm) and irregular grains of somewhat altered ilmenite (0.04 to 0.2 mm) are widely disseminated. The contact of this siltstone with underlying shale is sharp, though slightly undulating, and has small displacements along slip planes of cleavage which are about 0.08 mm apart.

FINE-GRAINED DOLOMITE: A section was cut from a 1-foot bed of dolomite that crops out on the south shore of St. Albans Point. This bed grades laterally within 4 feet into thin-cleaving shale, and has 1–5 mm laminae, though many other dolomite beds are dense and homogeneous. The thin section was estimated to be about 85 per cent dolomite, which forms a mosaic of predominantly equidimensional euhedra varying little in size, averaging about 0.036 mm. In the darker laminae there are irregular, fine, flaky masses of black, opaque carbonaceous material commonly associated with pyrite disseminated in extremely small grains (0.008 to 0.03 mm). Quartz grains are uncommon, very small (0.018 mm ±), euhedral, rounded, or irregular. Irregular and euhedral quartz grains commonly have minute inclusions with high birefringence, assumed to be remnants of replaced dolomite. Sericite flakes (0.005 to 0.036 mm) are rare. Rounded small grains (0.01 mm) of very high relief and extreme birefringence, probably zircon, occur rarely. A matrix of extremely small argillaceous flakes makes up about 10 per cent of the rock.

The fine-grained dolomite resembles the shale in composition except for its far greater proportion of dolomite, which seems to be the result of replacement of argillaceous material. Apparently siltstone beds were dolomitized because of greater proportion of detritus, and presumably greater porosity and permeability in the original sediment. Since the siltstone beds generally become finer-grained upward, one may assume either decreasing permeability upward or an increasing distance upward from the more freely circulating solutions providing the carbonate, or both. This doesn't explain the lateral gradation from dolomite into shale where there is no evidence of original variation in grain size, and hence no indicated original variation in porosity or permeability. There is no correlation of this rare lateral gradation with joints or faults.

Hathaway Formation

General description, submarine slide breccia.— The Hathaway formation, named for Hathaway Point, on southeastern St. Albans Point, St. Albans, Vermont, designates argillite and bedded radiolarian cherts, commonly intensely deformed, with included small fragments to large blocks of quartz sandstone, coarse graywacke, dolomite, limestone, and chert. Some fragments strongly resemble dolomite and dolomitic siltstone beds of the underlying Iberville, but the coarse sandstone, chert, and graywacke are unlike any strata in the Iberville. Where the Hathaway and Iberville are in contact or in close proximity, there is marked disparity in intensity and nature of their deformation. The Hathaway appears to have deformed by flowage without development of good cleavage, commonly with disintegration of less mobile beds into blocks and boulders. The Iberville has undergone much less intense folding and faulting of a type normally associated with the regional deformation. For these reasons the Hathaway is inferred to be a submarine slide breccia initially deformed while its muddy constituents were still soft. The Hathaway, first recognized by Hitchcock (1861, v. I, p. 310) on St. Albans Point, was included with his "Hudson River group." Gordon (1923, p. 200) compared these outcrops with "thrust inliers," presumably klippen.

Limited exposures of the Hathaway occur on Hathaway Point and on two knolls 1.6 miles N. 15° W., and 1.5 miles N. 26° W. from the northwest corner of St. Albans Bay. Similarly deformed argillite with included fragments of fine-grained dolomite, dolomitic siltstone, and black to gray chert crops out on the east shore of the lake for 1200 feet northward from Beans Point. Additional very limited outcrops occur on the east lake shore at the Chittenden-Franklin county line; 2000 feet southward from the intersection of the shore with the county line; and 100 feet inland 2200 feet northward from the county line.

The best exposures of the Hathaway are less accessible, on islands. Rock and Lazy Lady islands, east of Hathaway Point, are composed entirely of the Hathaway formation. It is found on northwestern Woods Island and northeastern Knight Island. Butler Island is underlain by Hathaway except for small areas of Iberville on the northwest shore.

Beans Point, west Milton.—The Hathaway outcrops on the east shore of the lake from Beans Point northward to the county line lie in the zone of intense deformation close below the thrust at the base of the Highgate Springs slice. Because the more northerly exposures are so deformed tectonically, only the Beans Point area allows profitable examination of the relationship of the argillite to the Iberville. The argillite is mottled olive gray (5Y4/1) to dark greenish gray (5GY4/1) to greenish black (5GY2/1). On a polished surface cut perpendicular to foliation the mottled colors are seen to represent original bedding which has been folded most intricately, and sheared with no development of slickensides or breccia. This small-scale shearing has completly healed. Some minute fold crests merge into the adjacent bed, a streaming of one bed into the next with no sharp boundary such as is found on the flanks of the small, tightly compressed folds. Included in the argillite are rounded fragments of moderate-yellowish-brown (10YR5/4) weathering, dark-gray (N3), fine-grained dolomite, and cross-laminated dolomitic siltstone, identical with the dolomitic beds of the Iberville. The fragments are 4 by 7 by 20 inches and smaller, subangular to rounded. The long axes of most of the boulders plunge about 55° S. 45° E. The axial planes of the complex folding approximately parallel a rough, irregular foliation, N. 25° E.–60° SE., and the fold axes plunge generally S. 43° E. Foliation causes the argillite to split into irregular tapered chips about 3 by 1½ by a quarter of an inch, apparently similar to the phacoidally cleaved mudstones with blurred or streaked bedding described by Jones from Denbighshire, North Wales (1937, p. 246), which he believes were involved in submarine slides.

Thirty-six feet of cover separates the north end of the outcrop of argillite at Beans Point from Iberville rock composed of about 95 per cent thin-cleaving shale in which half an inch to 2-inch laminated dolomitic siltstone beds occur at 4- to 8-inch intervals, and 4- to 8-inch fine-grained homogeneous dolomite beds occur about 20 feet apart. The Iberville beds are overturned, N. 35° W.–46° NE., with cleavage N. 15° E.–35° SE. In the Iberville there is clearly tectonic deformation of indurated rock. In the argillite the foliation, fold axial planes, and long axes of included fragments more or less parallel cleavage of the near-by Iberville, but the argillite beds when initially deformed must have been highly plastic, almost fluid, and the Iberville was not. Small-scale structures of the argillite bear no relation to larger-scale folding of the Iberville along the shore farther north, except that minor thrusting in the argillite shows drag and slickensides. Thus the argillite has been deformed twice, first presumably as a layered mud with a high percentage of water, on or closely below the sea bottom, and again after induration when the Iberville was folded. The first disturbance of the argillite may well have been the result of subaqueous sliding, which accounts for lack of deformation of the underlying Iberville at that time. The argillite appears to lie in a southward-plunging overturned syncline, above the Iberville beds, but tectonic deformation in this area is so intense that one cannot rule out the possibility that it lies in considerable thickness, interbedded in the Iberville.

Southern St. Albans Point, including Hathaway Point.—The Hathaway formation on southern St. Albans Point, including Hathaway Point, lies on the southeast flank of an anticline plunging 13° N. 50° E. The anticlinal

crest in the Iberville is at the southwestern tip of St. Albans Point. Across a beach about 250 feet east of the anticlinal outcrop, the Iberville is rumpled in tight folds with axial-plane cleavage N. 50° E.–72° SE. About 400 feet farther east, the Iberville is more intensely crumpled, with boudinage in the dolomitic beds. Bedding here parallels the cleavage, N. 37° E.–66° SE. On the shore northeastward, around a small point, 1200 feet N. 72° E. from the southernmost tip of St. Albans point, still younger beds appear with the attitude N. 32° E.–62° SE. They are 31 feet of light-gray to white weathering, grayish-black (N2) chert beds 2–6 inches thick, averaging about 4 inches. A thin section cut from an identical chert bed on Hathaway Point revealed Radiolaria in varying degrees of destruction. Overlying the chert beds is black siliceous argillite in which bedding is not readily apparent because of its irregular foliation similar to that at Beans Point. The argillite contains rounded pebbles (averaging 1 by 2 inches) of gray dolomite and fragments of chert. Some graptolites found in the argillite resemble *Dicranograptus*, *Dicellograptus*, and *Climacograptus*, but smearing from deformation precluded definite identification. Exposures on this shore line prove that the chert and disturbed argillite structurally overlie the Iberville beds at the southern tip of St. Albans Point.

In the middle of the beach 300 feet east of the above described chert and argillite, there is a small outcrop of Iberville with undulating minor folds plunging 5° N. 28° E. imposed upon a general southeasterly dip. About 400 feet east of this outcrop, across the rest of the beach, are the disturbed chert and argillite of Hathaway Point. The structural relationship of the mid-beach Iberville cannot be deter-

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PLATE 2.—DETAILS OF IBERVILLE BEDDING

FIGURE 1.—DETAIL OF RHYTHMIC BEDDING

Scour surface with overlying laminated dolomitic siltstone lies above eraser. Each cycle consists of dolomitic siltstone, dark-gray noncalcareous shale, and grayish-black noncalcareous shale. Exposure in rip-rap quarry 1.6 miles S. 10° E. of North Hero-Alburg bridge

FIGURE 2.—RHYTHMICALLY BEDDED SHALE

Noncalcareous shale and dolomitic siltstone (lighter beds) showing fine fracture cleavage, at Clark Point on Hog Island, west Swanton

FIGURE 3.—GRADED AND CURRENT-RIPPLED DOLOMITIC SILTSTONE

Lower part of picture shows a thick graded bed of dolomitic siltstone with angular shale fragments, and intrastratal crumpling toward the right. It is capped by laminated dolomitic siltstone with ripple drift. Ripple drift is better shown in the next higher siltstone bed. Exposure on the eastern tip of Burton Island

PLATE 3.—IBERVILLE BEDDING DETAILS AND STONY POINT TECTONIC STRUCTURES

FIGURE 1.—IBERVILLE RIPPLE DRIFT AND FLOW (LOAD) CAST STRUCTURES

At bottom, dolomitic siltstone with flecks of shale grading up into shale, showing crude bedding and penecontemporaneous slump slices. Upper laminated dolomitic siltstone has drift ripples on top and flow (load) cast structures on underside. Eastern tip of Burton Island

FIGURE 2.—IBERVILLE DOLOMITE GRADING LATERALLY TO SHALE

Fine-grained laminated dolomite (right of pick) grading laterally to finely cleaved shale (left). South shore of St. Albans Point

FIGURE 3.—MINOR THRUST IN THE STONY POINT

Minor thrust (tip of pick) parallels finely cleaved calcareous shale above, and cuts at a low angle coarsely cleaved argillaceous limestone below, where thick fracture cleavage plates have been rotated counterclockwise as shown by effects of light stratal bands below and left of pick handle. South shore of Colchester Point

FIGURE 4.—RUPTURE LENS IN STONY POINT

Fine-grained limestone rupture lens in calcareous shale and argillaceous limestone close below Champlain thrust. East shore of Lake Champlain 2.6 miles south of Beans Point

FIGURE 1

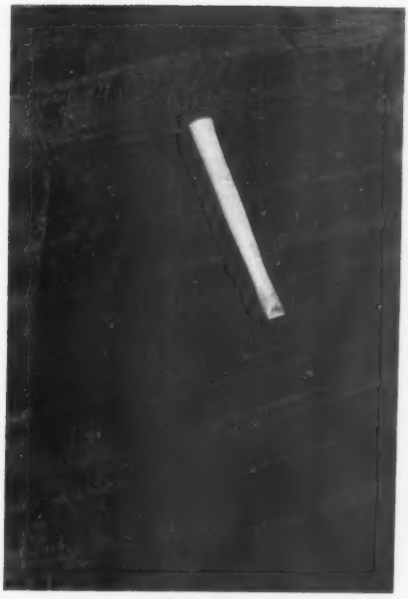

FIGURE 2

FIGURE 3

DETAILS OF IBERVILLE BEDDING

FIGURE 2

FIGURE 4

FIGURE 1

FIGURE 3

FIGURE 1

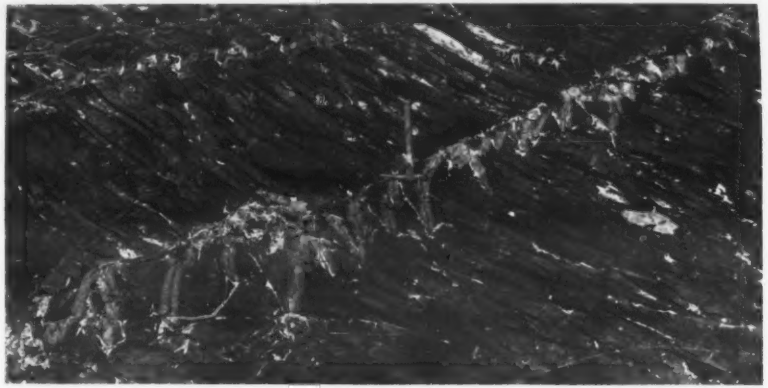

FIGURE 2

FIGURE 3

FIGURE 4

TECTONIC STRUCTURES IN STONY POINT AND IBERVILLE

FIGURE 2

FIGURE 4

FIGURE 1

FIGURE 3

DETAILS OF THE HATHAWAY

mined. It may represent (1) Iberville-type shale and dolomite interbedded in the lower Hathaway; (2) an outcrop of Iberville close to the crest of an eroded anticline covered by the beach to the west; (3) a repetition of the Iberville by thrusting, for there are many minor thrusts to the east on Hathaway Point. Probably the chert and argillite on Hathaway Point are younger than the Iberville outcrop in the middle of the beach.

On Hathaway Point the Hathaway formation is a pseudoconglomerate with a matrix of pale-greenish-yellow (10Y8/2) weathering rock seen on a polished surface to be composed of small, irregular, curdled masses of greenish-gray (5G6/1) to olive-gray (5Y4/1) argillite. Streamed and isoclinally folded in the matrix is black siliceous argillite similar to that in which the graptolites were found. In the intensely deformed matrix are small masses of grayish-black (N2) radiolarian chert which are commonly angular and in places attenuated parallel to bedding and foliation, as well as masses of bedded chert measurable in tens of feet. Fragments of dolomite and dolomitic siltstone occur only in the western portion of the Hathaway Point exposure, which appears to underlie structurally the central and eastern portions. Numerous slickensided tectonic shears, however, make it impossible to reconstruct reliably any original order of deposition. As shown on the map (Pl. 1), where only a representative sample was plotted, the shears and slickensides have a wide variety of orientations. One 40-foot wedge between shears is composed of isoclinally folded (axial planes N. 37° E.–75° SE.) calcareous shale (about 80 per cent) and noncalcareous shale (about 20 per cent), with occasional boudinaged masses of dense, fine-grained limestone about 2 inches thick. This lithologic suite resembles the transition beds at the base of the Iberville.

PLATE 4.—TECTONIC STRUCTURES IN STONY POINT AND IBERVILLE

FIGURE 1.—COMPRESSION EFFECTS IN THE STONY POINT

Buckling and slippage overlap of a 6-inch thick, fine-grained limestone bed in finely cleaved calcareous shale. Southwest Ball Island

FIGURE 2.—COMPRESSION EFFECTS IN THE STONY POINT

Zig-zag buckling of a 2-inch thick, fine-grained limestone bed in argillaceous limestone which has thickened by solid flow and formed coarse fracture cleavage. Bedding parallels the buckled limestone zone. East shore of Grand Isle northwest of north end of Savage Island

FIGURE 3.—BEDDING-PLANE SLICKENSIDES IN IBERVILLE

In interbedded calcareous and noncalcareous shale the crushed and slickensided shale has been replaced by calcite (white surface).

FIGURE 4.—STONY POINT CRUMPLED BELOW CHAMPLAIN THRUST

In laminated limestone (light) and calcareous shale (dark) close below the Champlain thrust, fold axes approximately parallel direction of thrust transport. East shore of Lake Champlain 1.9 miles south of Beans Point.

PLATE 5.—DETAILS OF THE HATHAWAY

FIGURE 1.—MASSIVE CHERT AND MASHED ARGILLITE

Crudely bedded chert (massive at top), argillite (streaked across center), and pseudoconglomerate with blocks of chert and dolomite (bottom). Northern point, west side of Woods Island

FIGURE 2.—PSEUDOCONGLOMERATE

A cobble of dolomite (lower right) and fragments of dolomite and black chert "float" in a flowage mash of argillite. Northern point, west side of Woods Island

FIGURE 3.—LIGHT AND DARK STREAKY ARGILLITE

Lower left is half of a silty dolomite block. Argillite shows streaming and swirling of original light- and dark-gray strata. Smaller black fragments are chert. Southwest point of Butler Island

FIGURE 4.—DOLOMITIC QUARTZ SANDSTONE BLOCK IN ARGILLITE

Mashed argillite wraps around the block of dolomitic quartz sandstone. Detail of small fragments and streaked matrix of the argillite is seen above the nickel. Southwest point of Butler Island

Polished surfaces of the argillite from Hathaway Point show that during one phase of its deformation it was almost fluid, perhaps a slurry of layered muds. Yet the same argillite occurs as cataclastic breccia along some of the minor faults. Probably the varied attitudes of shears were determined during submarine sliding, with renewed movement during tectonic deformation.

Islands in St. Albans Bay.—Two islands near the center of St. Albans Bay appear to lie along the axis of a complexly folded and sheared, northeastward-plunging major syncline. Rock Island, the smaller southern island east of Hathaway Point, is composed of greenish-gray argillite with rough foliation N. 27° E.–55° SE.; it contains occasional rounded to angular fragments of dolomite and dolomitic siltstone as large as 3 by 8 inches. Lazy Lady Island has rusty-weathering black argillite and black to gray bedded chert, having an attitude of N. 65° E.– 35° SE.

Woods Island, west St. Albans.—A crucial outcrop of Hathaway chert and argillite on the small point on the west shore of Woods Island 1400 feet southward from the northern tip of the island is crudely bedded (Pl. 5, fig. 1). The gross structure is a shallow, asymmetrical syncline plunging 17° N. 55° E. The southeast limb is steeper than the northwest, conforming with the regional tectonic pattern. At the bottom of the section an 8- to 10-foot layer of foliated pseudoconglomerate of streaked and crumpled lighter and darker gray argillite warps around a great variety of included rock fragments (Pl. 5, fig. 2). These fragments are: (1) irregularly angular to well-rounded blocks and cobbles of dolomite and laminated dolomitic siltstone from half an inch by 3 inches to 18 by 24 inches; (2) small rounded pebbles (1 inch ±) of black chert; (3) thin fragments (a quarter of an inch by 12 inches to 1 by 36 inches) of beds of coarse calcarenite rich in disseminated fine pyrite, composed chiefly of broken pieces of brachiopods, gastropods, and crinoid stems, some of which have been replaced by pyrite, and angular chips of noncalcareous shale. The shelly calcarenite seems to have originated in fairly shallow water. Above the 10-foot zone of pseudoconglomerate is a 10-inch bed of dark-greenish-gray (5GY4/1) argillite which is rumpled; it thickens and thins to form incomplete boudins. A polished section of this argillite shows fine black streaks which apprently originally defined bedding and are now displaced as much as 2 mm along irregular surfaces. The streaks thin, branch, and swirl irregularly into the matrix. Fine, disseminated pyrite is common. At the top of this outcrop is about 5 feet of massive, sheared, and rumpled dark-gray (N3) to dark-greenish-gray (5GY4/1) chert. A polished surface of the dark-gray chert reveals small angular blocks of chert showing bedding traces, in a swirled matrix of dark-gray and medium-dark-gray chert with fine disseminated pyrite. The matrix has healed these ruptures into a dense hard rock which breaks under the hammer in total disregard for internal structure. The rock was only semiindurated when originally disturbed, possibly jell-like. Northward from the point for 400 feet along the shore are intermittent outcrops of argillite similar to the lowest 10-foot layer described above, but with included blocks only of dolomite and dolomitic siltstone up to 6 by 1½ feet. Four hundred feet north of the point, indefinite bedding is approximately N. 52° E.–22° NW., suggesting that the massive chert on the point underlies this rock, and that the section of disturbed beds is fairly thick.

Structural relationship of the Hathaway on the point, and the measured sections of the Iberville farther south on the west side of Woods Island indicate that Hathaway overlies Iberville. Throughout the Iberville sections the bedding dips northwest, except for overturned beds at the southern end. The northwest dip gradually decreases northward along the shore toward the place where the uppermost beds of the section lie N. 49° E.– 30° NW. Thus, the measured sections of Iberville are on the east flank of a large syncline gently plunging slightly east of north. Separated from the northwest end of the section by 90 feet of beach is an isolated outcrop of Iberville in which a small gentle anticline reverses the dip direction for a few feet. In another outcrop, 400 feet northwest of the measured section, typical Iberville beds lie N.-S.–18° E. At the northwest end of this outcrop the beds are broken by weathering and partially covered by shingle, but appear to resume a northwestward dip. This is 100

feet southeast of the Hathaway outcrop on the point.

Knight Island, east North Hero.—The structural relationship of the Hathaway, in the northeastern corner of Knight Island, to the Iberville on the rest of the island also indicates that Hathaway is younger than Iberville. Where Iberville is exposed, displacement on a few minor thrust faults is insufficient to juxtapose unlike rock types. Drag folds along the southwest half of the southeast shore range from a large (wave length 20 feet, amplitude 8 feet) asymmetrical fold plunging 15° N. 55° E. to small tight overturned folds plunging 14° S. 5° E. Axial planes of these minor folds dip eastward, and throughout the island the prevailing southeastward dip of the beds is gentler than the southeastward-dipping cleavage, suggesting that the island lies on the western limb of a syncline. On the southeast shore about 2100 feet southwestward from the easternmost tip of the island is a fault zone where the rocks of the northeastern end of the island have moved downward, and probably northwestward, relative to the Iberville southwest of these faults. Either or both of these movements, considering the assumed syncline to the east, would emplace younger rocks on the northeast against the Iberville. The rocks northeast of these faults are entirely of the Hathaway type.

The Hathaway formation of northeastern Knight Island resembles that of northwest Woods Island and Hathaway Point in its mashed gray to greenish-gray argillite, with included boulders (from 1 by 2 inches to 1 by 3 feet) of fine-grained dolomite, laminated dolomitic siltstone, and fragments of chert. Fine- to coarse-grained graywacke boulders as large as 5 feet also occur on Knight Island. The attitude of the foliation in the argillite varies from N. 17° W.–45° NE. to N. 33° E.–65° SE. About half way between the fault zone and the easternmost tip of Knight Island, 15 feet of bedded grayish-black (N2) to dark-gray (N3) chert lies N. 25°W.–50° NE., parallel to argillite foliation in the immediate vicinity. On the eastern point of Knight Island 8 feet of dark-gray bedded chert, in beds of 2 to 8 inches, lies N. 18° E.–47° SE., between zones of mashed argillite, one of which includes a boulder of dark-gray coarse-grained graywacke (4 by 8 feet). Immediately east of the chert beds a 40-foot zone of calcareous shale and argillaceous limestone, with a dense fine-grained limestone bed 10 inches thick, tapers toward the south. One 6-inch bed of shale rich in pyrite has weathered a rusty brown. A 1-inch bed of shell-fragment calcarenite is similar to that on Woods Island. Traced southward laterally along the bedding this wedge of rock disappears into the argillite with no slickensides or cataclastic brecciation to suggest tectonic imbrication. Bedding in the wedge lies N. 36° E.–54° SE. It was probably emplaced during submarine sliding.

Butler Island, east North Hero.—Almost all of Butler Island is underlain by the Hathway disturbed argillite and chert. Limited exposures of Iberville crop out on the west shore 2000 to 4000 feet northward from the southwest corner of the island, striking N. 22°–40° E., and dipping 65°–70° SE. Cross-lamination in the dolomitic siltstone interbeds indicates that they have not been overturned and that the rocks southeastward from this shore are probably younger. About 2100 feet northward from the southwest corner of the island is a thrust fault, N. 25° E.–20° SE., with slickensides indicating that the hanging wall moved N. 37° W. relative to the footwall. Typical Iberville underlies the thrust, below a 2-foot breccia zone. Overlying is Hathaway mashed argillite with inclusions of dolomite, dolomitic siltstone, and occasionally black chert, from 1 by 2 to 8 by 24 inches. Inasmuch as the Iberville strata dip more steeply southeastward than the fault, this probably represents thrusting of younger rocks over older. In the argillite between the fault and the southwest corner of the island are several large boulders of medium-dark-gray (N4) weathering, dark-gray (N3) graywacke from 1 by 2 to 5½ by 11½ feet.

The Hathaway of the southern half of Butler Island has been sheared with many minor low-angle thrusts, dipping southeastward to eastward and cutting across the chaotic folding and foliation of the argillite. They are frequently associated with thin cataclastic breccia zones and with slickensides which indicate the hanging walls moved westward to north-

westward relative to the footwalls. They are definitely a part of the tectonic pattern of deformation but do not seem to have any significant displacement. Intricate folding of the argillite, with occasional limestone, argillaceous limestone, and calcareous shale beds, approaches the pattern of ptygmatic folding in some places and seems, in part at least, to be the result of deformation of soft sediment.

On the southwest corner of Butler Island the argillite is similarly tightly compressed into ruptured isoclinal folds (axial planes and foliation: N. 19° E.– 64° SE.), with extensive flowage, and it includes pebbles, blocks, and boulders of chert, dolomite, dolomitic siltstone, fine-grained limestone, and graywacke (Pl. 5, figs. 3, 4). Between the southwest corner and the southernmost tip of the island, distinct bedding is apparent in calcareous shale and cross-laminated calcareous quartz siltstone. These beds are in the core of an overturned isoclinal anticline about 50 feet across. On the flanks is noncalcareous shale with occasional 2- to 6-inch dolomite beds, similar to the Iberville. The smaller-scale deformation involves rupture and displacement of the more brittle beds by flowage of the argillaceous beds. Dolomite beds as thick as 7 inches are tightly folded without development of fracture cleavage which one would normally expect if these beds were well indurated at the time of deformation. These folds plunge about 43° S. 1° E., on an axial plane having the attitude N. 25° E.–65° E. The more competent beds have commonly sustained rupture and boudinage. On the east flank of the anticline is a slice of intercalated limestone and dolomite (2–4 inches) and shale beds (1–2 inches) about 5 feet thick in angular discordance with the overlying and underlying beds of similar rock. Bounding surfaces of this slice are sharp, with no slickensides or brecciation. These beds dip steeply southeastward almost parallel with overlying bedding. Emplacement of this slice does not seem to have been contemporaneous with brecciation and slickensided shearing allied to the regional tectonic pattern, and movement before complete lithification is inferred.

East of the anticline the mashed argillite with its included boulders and fragments is encountered again. Close association of bedded rock in the anticline, and the mashed argillite east and west of it, suggests that the argillite must have been almost fluid when it was initially deformed, apparently by submarine landslide, and that the bedded rocks underlying were somewhat more competent, though not so well indurated as they are today. The discordant emplacement of the bedded slice suggests that it, too, was involved in the slumping or sliding. Baldry (1938, p. 353) described similar phenomena interpreted as resulting from large-scale submarine slip, in the Eocene and Oligocene of Peru, where "alternating thin flags and shales are contorted in a kind of isoclinal folding. With thicker flags and less shale, the formations break into shear lenses of all sizes." Baldry, however, found minor breccia zones associated with slip planes. Where catalcastic breccia zones are present in the shale and argillite of the Champlain Valley, they are associated with tectonic faulting. Along the east mainland lake shore similar beds are plastically deformed by orogenic forces. On southeast Butler Island, only the absence of slickensides and brecciation suggests that the rocks in the anticline were deformed before deep burial and hardening. Admittedly, subsequent tectonic deformation has taken place, and it is difficult to determine the origin of some structures, whether tectonic or contemporaneous. Kuenen (1948, p. 379) reports similar structures in the Carboniferous of Pembrokeshire, south Wales, where lack of slickensides and crush phenomena casts some doubt upon the otherwise obvious tectonic origin.

From the southern point of Butler Island northeastward for 1¼ miles along the east shore, foliation and bedding of argillite with occasional chert beds up to 2 feet thick strike N. 20°–42° E., and dip 53°–80° SE. Bedding fragments of cross-laminated dolomitic siltstone are very common, along with blocks and fragments of chert, dolomite, quartzite, graywacke, and occasionally limestone. Most of these are aligned with their long axes parallel to the foliation. Since the foliation seems to be tightly compressed bedding parallel to rough cleavage, one can determine whether the layers are overturned only by the truncated surfaces of cross-laminae in the dolomitic siltstone.

Where such evidence in present in the three-quarters of a mile northward from the southern point of the island, the bedding has not been overturned. Along the shore for the next quarter of a mile, north of the rounded point, laminae indicate definite overturning, but beyond that, with no indication of a change in general attitude, the bedding fragments dip equally steeply east-southeast without overturning. Thus, if these fragments are reliable indicators of general overturning, the argillite is tightly compressed in isoclinal folds. Two blocks of dolomitic siltstone measured 2½ by 8 feet and 3 by 20 feet. Three blocks of coarse-grained graywacke measured 3 by 8 feet, 6 by 10 feet, and 15 by 50 feet. Argillite foliation wraps around these blocks and around innumerable smaller pebbles and boulders.

Along the southern shore of the easternmost promontory of Butler Island isoclinally folded argillite contains many large blocks similar to those described from the southeast shore of the island. Bedding and foliation strike N. 27°–53° E. and dip 49° SE. (normal) to 29° SE. (overturned). Around the point on the east shore of this promontory structural trend and rock type change markedly. Though intensely deformed, the rock resembles Iberville through the central portion of the zone of deformation extending for 1300 feet northward from the southeast corner of the promontory. It is flanked on the north and south by similarly deformed beds of chert and argillite. Bedding strikes N. 35° W. to S. 88° W. and dips 70° NE. (overturned) to 45° NE. (normal). Isoclinal folding is common. One fold axis plunges 33° N. 60° W. with a vertical axial plane. Yet the thin cleavage in the shale (N. 36° E.–52° SE.) reflects the regional tectonic pattern! Thus, the mudstone had first been compressed in a direction perpendicular to later stress which produced its cleavage and the major regional thrusts and folds. Since apparently no cleavage resulted from the earlier deformation, one may assume that the rock was not sufficiently indurated to develop cleavage.

The northeastern shore and the northern end of Butler Island have a chaotic structural pattern, with alternating areas of mashed argillite and chert, and of Iberville-type beds. Bedding trends are generally northwest–southeast, and in many places the beds are vertical or overturned toward the southwest. Fine cleavage, where developed, adheres to the regional northeast–southwest strike.

On the northern shore of the island, directly below argillite containing a 4-foot bed of black chert, a breccia of limestone blocks (up to 2 by 2½ feet) in a matrix of calcareous shale extends for about 600 feet along the shore. Included in the breccia are blocks of medium-light-gray (N6) calcilutite, very fossiliferous medium-dark-gray (N4) coarse calcarenite, dark-gray (N3) calcilutite, and dark-gray (N3) argillaceous limestone. A vertical 6-inch bed of fossiliferous conglomerate, striking N. 45° W., contains limestone pebbles that are occasionally rounded but more commonly angular, as much as 1 inch in diameter, embedded in a matrix of fine limestone fragments and calcareous argillaceous sediment.

In two of the large fossiliferous calcarenite boulders, the following were identified:

Rafinesquina sp.
Resserella sp.
Sowerbyella sp.
Platystrophia sp.
Cryptolithus tesselatus Green
Streptelasma corniculum Hall
Dalmanella rogata Sardeson

This faunule is characteristic of lower Trenton, and the assemblage is common in the Shoreham member of the Sherman Fall formation.

Various boulders may represent several formations older than the Hathaway: Shoreham Cumberland Head, and Stony Point. The jumble of rock types, the chaotic structure, and the close association with Hathaway chert and argillite suggest that these older rocks, too, were involved in submarine slides. If so however, it is evidence of a deeper disturbance than elsewhere, possibly even large mass movements off a submarine fault scarp, similar to those in Scotland (Bailey and Weir, 1933, p. 457), and in Quebec (Bailey, Collet and Field, 1928, p. 603).

One must consider that this breccia might be the result of the normal faulting so common in the Champlain Valley. Cady (1945, p. 571) summarized the evidence indicating that Adirondack normal faulting postdates thrusting in west-central Vermont and is late Ordovician

or Silurian (1945, p. 581). Many minor tectonic shears in these rocks produced extensive slickensiding and minor cataclastic brecciation, which seem contemporaneous with major thrusting because they correlate with attitudes of shears and bearings of slickensides associated with the thrusts. They indicate considerable induration at the time of thrusting. Thus, if normal faulting came later than thrusting, it could hardly have resulted in a fault breccia of large blocks of well-indurated Shoreham and Cumberland Head dragged upward along a fault zone past 1000 feet of Stony Point, and at least another 1000 feet of Iberville. An alternative explanation is that a normal fault older than the thrusts produced a scrap from which this material slumped, in which case one would expect to find Shoreham and older rocks cropping out in the vicinity of this conglomerate, probably to the north. Unfortunately, the bedrock lies below the lake for many miles to the north. No evidence of normal faulting of such magnitude has been found in this area, except possibly in South Hero, and that appears to be related to later faulting found on the flanks of the Adirondacks. Tentative elimination of normal fault scarps to account for the submarine sliding in the Hathaway leaves the possibility of thrust scarps. Present outcrops of the thrusts at the bases of the Highgate Springs and Rosenberg slices are only 4¾ and 5½ miles, respectively, east of Butler Island. At Lime Rock Point, on the east shore of St. Albans Bay, and on the north westernmost of the two bedrock knolls about three-quarters of a mile northwest of St. Albans Bay Village (at the northeast corner of St. Albans Bay), massive Canadian limestone of the Highgate Springs slice lies on crushed and rumpled Iberville. The submarine sliding evidenced in the Hathaway may have originated in early movements along these thrusts, but only a fault scarp of considerable magnitude could have provided boulders from strata at least 2000 feet conformably beneath the matrix in which they lie on the north end of Butler Island.

Thin sections of Hathaway.—Thin sections were cut from radiolarian chert and argillite beds, a sandstone boulder, and a graywacke boulder from the Hathaway formation.

RADIOLARIAN CHERT: Radiolarian chert was sectioned from bedded chert from the eastern tip of Hathaway Point. The rock is moderate-yellowish-brown (10YR5/4) to very-light-gray (N8) weathering, dark-gray (N3) chert. Radiolaria appear as: (1) brown, roughly circular areas 0.145 mm (±) across, with a rim of very small opaline grains and an interior almost entirely isotropic, with occasional faint gray flecks showing through crossed nicols; (2) clear, roughly circular areas similar to (1) in size and shape, but filled with fine-grained chalcedony. Some of these, better preserved, have fragmental radial spines still in place on the perimeter, but none show clearly the internal structure. Most of the Radiolaria are deformed or crushed. Sixty per cent or more of the rock consists of these tests and chalcedonic and opaline framgents of spines and tests about 0.01 mm across. Most fragments are about equidimensional, but occasional fractured branching and cruciform spicules are 0.49 mm long. Similar thin sections have been interpreted by King (1937), Henbest (1936), and Ruedemann and Wilson (1936) as remains of Radiolaria and sponge spicules. The groundmass is isotropic to faintly birefringent, amorphous-appearing material. Occasional angular quartz fragments average about 0.004 mm across, but some are as large as 0.036 mm. One rounded grain of quartz, 0.08 mm across, has a narrow rim of secondary chalcedony. Extremely small (0.0009 mm ±) sericite flakes are common; a few are as large as 0.018 mm. They, with the more elongate fragments of quartz, show rough lineation, but since bedding was indeterminable it is not know what the lineation represents.

ARGILLITE: Argillite sectioned perpendicular to the foliation from dark-olive-brown (5Y4/2) weathering, medium-gray (N4) argillite on the southwest corner of Butler Island is composed predominantly of cryptocrystalline clay minerals with many minute isotropic grains of high relief, possibly opaline fragments of Radiolaria. A few roughly circular isotropic areas (0.03 mm ±) may be remains of radiolarian tests. Sericite flakes (0.001 by 0.004 mH ±), densely scattered through the groundmass show two general directions of lineation, 35° apart in the plane of the section. Quartz occurs here and there, as angular fragments (0.007 0.018 mm ±) or rounded grains (0.018 mm ±). Pyrite is very common and is disseminated in small grains (0.008 mm and smaller).

SANDSTONE BOULDER: A large boulder of dark-gray (N4) well-indurated sandstone lies in the argillite on the west flank of the anticline on the south shore of Butler Island. It is about 70 per cent well-rounded quartz grains, commonly with secondary enlargement, averaging about 0.21 mm across, though a few are 0.40 mm. Plagioclase grains with polysynthetic twinning are very common; many are somewhat decomposed and smaller than the quartz

(0.06 to 0.14 mm). Symmetrical extinctions on twin lamellae were measured on four grains: 11° and 12°, 14° and 15°, 14° and 16°, 20° and 20°. These would correspond to oligoclase and either albite or andesine, as did the feldspar of the graywacke from northeastern Knight Island. Carbonate filling of intergranular space is common, and intergrown with it is yellowish-brown, translucent to opaque material believed to be limonite. Pyrite is rare, in occasional cubes (0.065 mm ±) and anhedra (0.16 mm ±). Biotite flakes are rare, broken and bent between the other detrital grains. Heavy-mineral grains are fairly common, ranging from 0.02 to 0.117 mm; zircon is the most abundant. One oval tourmaline grain (0.1 mm) was noted. Heavy minerals are concentrated along thin laminae with the pyrite and with smaller than average detrital grains of quartz and feldspar. Such detrital mineral assemblage could have resulted from erosion of either a metamorphic or igneous terrane, or both.

GRAYWACKE BOULDER: In an 8- by 36-inch boulder of medium-dark-gray (N4) graywacke in argillite on the east shore of Butler Island, about 4850 feet N. 28° E. from the southern point, the most common constituents are quartz and feldspar. Some of the larger quartz grains (as large as 0.95 mm) are rounded, but most are angular and smaller (0.033 to 0.36 mm). Most quartz shows strain shadows, and some has fine sericite along cracks. Zircon and acicular rutile inclusions in the quartz are fairly common. Polysynthetically twinned feldspar is occasionally rounded, but more commonly angular, ranging from 0.08 to 0.36 mm. Much of the feldspar is partially decomposed to carbonate. Symmetrical extinction measurements on twin lamellae of two grains were: 11° and 12°, and 12° and 13°, indicating oligoclase bordering on either albite or andesine. Some microperthite grains (0.3 mm ±) were noted. Finely disseminated chloritic material is very common; some of the occasional large grains (0.2 mm ±) have inclusions of limonite and carbonate. Biotite flakes (0.2 to 0.08 mm) were occasionally found, and, somewhat more often, muscovite (0.35 by 0.02 mm), bent and broken between the other detrital grains. Carbonate occurs in angular grains (0.26 by 0.41 mm ±), and as irregular growths (0.05 to 0.2 mm ±). Of the rarer heavy minerals, rounded zircons are fairly common (0.065 to 0.2 mm). Dark red-brown isotropic grains (0.02 to 0.15mm) of high relief, with a metallic reddish-gray luster by reflected light, are probably hematite. There are occasional small opaque white grains of leucoxene. Pyrite is comparatively rare, in anhedra up to 0.5 mm. Many brown opaque grains, commonly rounded (0.16 by 0.33 mm ±), seem to represent limonitic alteration of ferromagnesian minerals. A few small pleochroic blue-green grains 'up to 0.025 by 0.098 mm) appear to be fibrous actinolite. Rock aggregates of several types are common. There are one large (1.5 by 0.16 mm) and many smaller grains of shale. Slate, distinctive for lineation of fine micaceous flakes, with finely divided pyrite, occurs in rounded grains (0.16 by 0.33 mm ±). Quartz aggregates were found as large as 0.77 mm, both rounded and angular. A few particles (up to 0.98 by 0.42 mm) of a somewhat decomposed rock, made up of quartz, feldspar, and chlorite, may be of volcanic origin, as their structure resembles a porphyry. All this detritus is set in an extremely fine matrix of chloritic and argillaceous material.

STRUCTURAL GEOLOGY

Major Overthrusts

Along the Champlain thrust, lower Cambrian Dunham dolomite (Clark, 1934, p. 9; Cady, 1945, p. 528) overlies shales of the central lowland of the Champlain Valley northward through the field area to a point a quarter of a mile east of the lake shore in northwestern Milton Township 1¾ miles south of the Chittenden-Franklin county line. From here northward for 4 miles the Dunham is underlain by a very light gray weathering, finely crystalline limestone with buff-weathering dolomitic impurities, the Beldens formation (upper Canadian) of the Highgate Springs slice (Kay, 1950, p. 1476). Nowhere is this belt of limestone more than a third of a mile wide, and it disappears on the hillside about 2 miles north of the county line. The limestone is folded in open asymmetrical folds striking generally northeast–southwest. Northwestward dips are as steep as 56°, and the southeastward as steep as 32°. These folds plunge gently northeastward; one small anticline plunges 12° N. 35° E. Contact with underlying shale along the lake shore is nowhere exposed, but the shale is intensely crumpled and sheared, just as it is farther south directly beneath the Champlain thrust.

Autochthonous shales directly underlie the Champlain thrust from the northern end of this Beldens belt of outcrop northward to the base of the Champlain thrust escarpment a mile S. 18° E. from Lime Rock Point. Iberville float extends to the Dunham cliffs, with no evidence of the Beldens. Southeast of Lime Rock Point,

the Beldens appears in an anticline trending about N. 60° W. with a wave length of 75+ yards. At Lime Rock Point the Beldens is clearly thrust over the Iberville; this is the Highgate Springs thrust. The thrust surface here strikes N. 33° W. and dips 14° NE. On this surface fluting and slickensides are barely discernible, though a suggestion of fluting strikes N. 75° W.

The Highgate Springs thrust apparently continues southeast from Lime Rock Point with unchanged attitude, and is cut off by the overriding Champlain thrust (Pl. 1). Northward from Lime Rock Point, under St. Albans Bay, the Highgate Springs thrust must strike north–northeast. It appears again, with a local attitude of N. 15° W.–35° E., on the northwestern of two bedrock knolls three-quarters of a mile northwest of St. Albans Bay village, at the northeast corner of St. Albans Bay. The Highgate Springs slice, continuous northward to Missisquoi Bay, has been studied by Gordon (1923), McGerrigle (1931), and Kay (1945). Gordon described the thrust relationship in the old Swanton Lime Works quarry at Swanton, Vermont (1923, p. 192).

This study sustains Kay's interpretation that the Highgate Springs sequence is a thrust slice (1945, p. 1173) and defines its southern limit 1¾ miles south of the Chittenden-Franklin county line.

Deformation of Shales

Major folding.—Shales are extensively folded throughout the field area. Southward from southern South Hero folds strike generally N. to N. 12° E., but farther north, in the latitudes of North Hero and Alburg, the fold trends swing to N. 20°–30° E. Thus, they reflect the change in trend in the major folded mountain belt and belt of thrusts extending from eastern Quebec southward through the Green Mountains of Vermont (Cady, 1945, p. 562; Keith, 1923, p. 314; 1932, p. 363; Booth, 1950, p. 1167). As Gordon noticed in North Hero (1923, p. 163), there are north- to northeast-trending belts of tightly compressed and sheared folds, separated by broad areas of more gentle folding. Smaller folds on the larger folds plunge both northward and southward. The pattern of structural elements and formational boundaries indicates the northeastward plunge is more prevalent and perhaps a little steeper. The area might be visualized as having northeastward-trending folds imposed on an eastward regional dip, though there are many individual exceptions to this general picture.

The cumulative effect of innumerable smaller folds is difficult to evaluate in determining locations of major structures. Using the established lithologic sequence, however, in conjunction with bedding attitudes, several larger complex anticlines and synclines become apparent (Pl. 1). A syncline, in the core of which the Hathaway formation is found, trends north–northeast through the middle of St. Albans Bay and is covered at the northeast end by the Highgate Springs slice. Projected south–southwest, this syncline lines up with the north–northeast-plunging syncline in the core of which Iberville is found on northeastern Savage Island. Parallel to this syncline and on its northwest flank a complex anticline extends from Ball Island through Burton Island and southern St. Albans Point. A parallel syncline to the northwest, through Lapans Bay, is suggested by isolated Hathaway outcrops 1.5 miles north–northwest from the northwest corner of St. Albans Bay, by the northwestward dips on western St. Albans Point, and the eastward dip of Iberville beds north of Lapans Bay. Woods Island lies on the western overturned flank of an anticline separating the Lapans Bay syncline from a much broader, deeper syncline including the northwest part of Woods Island, Butler Island, and northeastern Knight Island. This Butler Island syncline must be either doubly plunging (north–northeast and south–southwest) or cut off by a fault between northern Butler Island and the lake shore northeast of there, for on that shore one finds only tightly folded Iberville beds. A gently north–northeastward-plunging anticlinal crest underlies the eastern side of North Hero Island, east and northeast of Cary Bay, and disappears northeastward into the lake. It loses its identity southwestward in the more gently folded Stony Point beds of southern North Hero. West of the overturned western limb of this anticline is a complex syncline in which a belt of Iberville extends from southwest of Cary Bay through northern

North Hero and Hog Island, west Swanton. Along the Alburg east shore, west of North Hero, is the crest of an overturned anticline which extends north–northeastward into the lake. The rest of Alburg is underlain by a series of more gentle folds.

The shale area of Grand Isle, except for the complexly folded and sheared east shore, is underlain by very gentle folds. Along the west shore gentle northward dip is almost constant for several miles. Eastern South Hero is more intensely folded. The west shore of Kibbie Point coincides with the western, vertical flank of an anticline. Parallel, and about half a mile east of this shore, is a deep, northeastward-plunging syncline in which Iberville beds are exposed. The east limb of this syncline is downfaulted against the western overturned limb of an anticline on which the Stony Point formation is exposed. Although covered, the fault is presumed to be a thrust because it parallels the trend of strongly asymmetrical folds, many of which are cut by minor thrusts. Elsewhere in eastern South Hero, folding is complex, but because of the comparative uniformity of the Stony Point shale no major folds can be distinguished.

In all larger tectonic folds, the northwest limbs are steeper and are commonly overturned.

Major faulting.—Although innumerable faults cut the shales, only a few displace them enough to juxtapose different formations. On most faults the rock of both walls is so similar that only minor displacement may be assumed. Block faulting typical of the western and southern Champlain Valley (Hudson, 1923; 1931; Quinn, 1933; Rodgers, 1937, p. 1583; Buddington and Whitcomb, 1941, p. 89–104; Cady, 1945, p. 570–572) is distinct only in the older Trenton, Chazy, and Canadian formations of western South Hero, where it has been studied by Kay and his students (Personal communication). Although Hudson (1931, Fig. 1) postulated many such faults through the shale areas, I found no direct evidence for them; all areal stratigraphic repetitions, omissions, or displacements can be explained by complex folding and minor ruptures. This does not exclude the possibility of larger-scale normal faulting under the lake.

Megathlin (1938, p. 106) suggested that the normal faulting in the Mohawk Valley of New York disappears "both by actual decreases in throw, and by passage into shale which is probably taking up some of the displacements by adjustments within itself." Cady (1945, p. 572) cited Megathlin with reference to rocks farther south in the Champlain Valley, and possibly some of the separated belts of more intense deformation (folding and shearing) in the shales of North and South Hero islands represent loci of normal faults below, in the more competent limestone.

Flow structures.—Relative competency of various kinds of beds in the Stony Point and Iberville formations is reflected in the nature of their deformation. Solid flow (Fairbairn, 1949, p. 4) has taken place extensively in the more argillaceous beds. The hard, purer limestone and dolomite beds have commonly buckled, ruptured, and overlapped, while the shale on either side has shortened and thickened by solid flow. This is shown on southern Ball Island (Pl. 4, fig. 1) where, in vertical beds of banded calcareous shale in the Upper Stony Point, a 6-inch bed of hard, fine-grained limestone buckled and overlapped by slip. On a point on the east shore of Grand Isle, 1.4 miles N. 20° W. from the northwest tip of Savage Island, a similar 2-inch limestone bed has undergone zig-zag buckling between thick beds of argillaceous limestone (Pl. 4, fig. 2). Implications of such deformation in measurements of stratigraphic sections have been discussed by Gair (1950, p. 872). Stratigraphic measurements of Iberville and Stony Point beds were made on the least deformed sections available, but some such thickening may have occurred in the Iberville of Woods Island and Clark Point.

Minor faults.—Shear along bedding planes, cleavage planes, and at varying angles to both is very common. In the more intensely folded belts, multiple shears occur along crests and troughs of folds where wedges of rock have been displaced upward out of folds or downward into them. The more common movements have taken place along low-angle surfaces, both of thrusting and of low-angle normal faulting (the hanging walls moved downward relative to the footwalls). In some places slickensides indicate low-angle strike-slip or oblique-slip faults.

Bedding thrusts are so common that the cumulative movement along them must be fairly great even though no measure of it is possible. Flexure folds occur in the more competent argillaceous limestone in areas of gentle folding. On the flanks of such folds bedding-plane slip should occur and might produce slickensides. However, bedding-plane slickensides extend across the crests of folds and thus must result from bedding thrust faulting. Typical bedding-plane slickensides, in the Iberville interbedded calcareous and noncalcareous shale on the southeast shore of Burton Island, are shown in Figure 3 of Plate 4. Frequently thrusts parallel bedding for 100 feet or more across the strike before cutting across the beds at a low angle, as on the south shore of Colchester Point (Pl. 3, fig. 3).

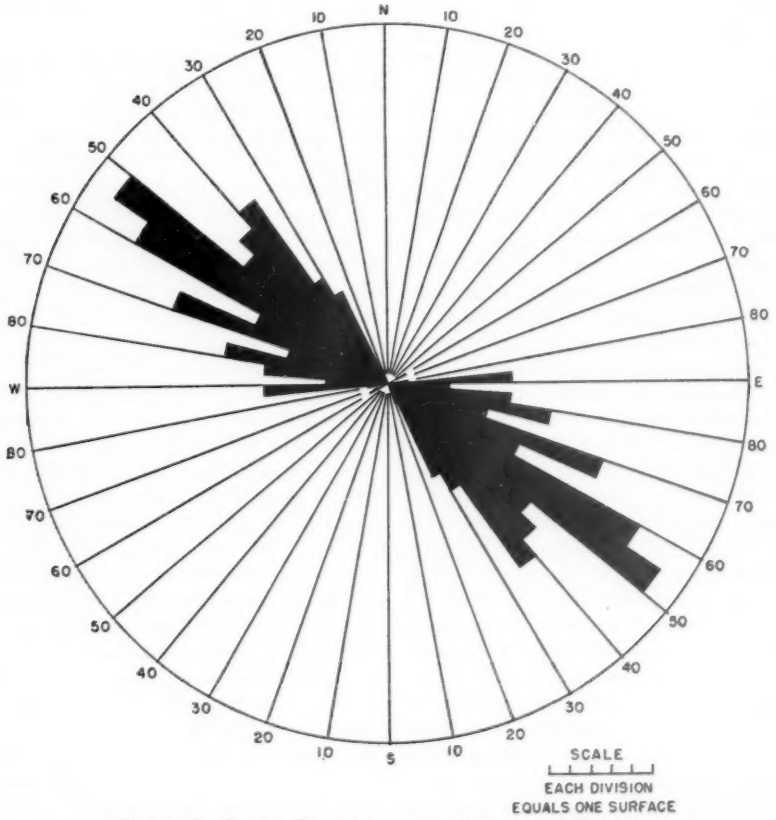

FIGURE 2.—RADIAL DIAGRAM SHOWING SLICKENSIDE BEARINGS

119 bearings of horizontal projections of slickensides on bedding and other low-angle shear surfaces plotted in 5-degree intervals.

Slickensides.—The bearing of slickensides is remarkably constant, regardless of the type of surface or its attitude. Slickensides everywhere indicate that the upper surface moved northwestward relative to the lower. One hundred nineteen bearings of horizontal projections of slickensides were plotted on a radial diagram (Fig. 2). Only three of these bearings lie outside the 70 degree arc between N. 25° W. and S. 85° W. Fifty-nine per cent of the bearings lie in the 25 degree arc between N. 35° W. and N. 60° W. To bring out any possible correlation

between the attitude of the shear surfaces and bearings of slickensides on them, the poles of 103 surfaces were plotted on the lower hemisphere of a Schmidt stereographic net (Fig. 3).

N. 40° W. (4 of them N. 40° W.), and 3 were N. 55° W.

Movement along multiple slickenside zones must have occurred in different directions at

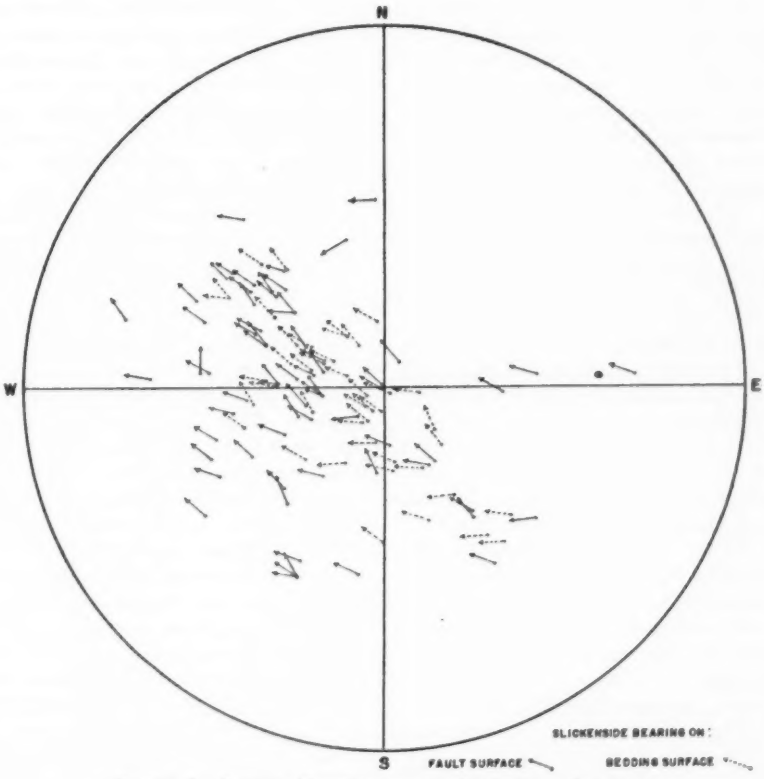

FIGURE 3.—SCHMIDT NET DIAGRAM OF SLICKENSIDE SURFACES

Poles of 103 slickensided surfaces are plotted on the lower hemisphere of a Schmidt stereographic net. For each surface pole, the compass bearing of its slickensides is plotted as an arrow. More than one arrow is drawn from the pole of a shear zone which has more than one slickensided surface with a different bearing for each surface.

There seems to be no correlation. Omitted in the stereonet plot were six curved surfaces and two surfaces with 15°–20° variations in bearing. Fourteen other bearings, plotted individually on the radial diagram, were combined on six surfaces on the stereonet because they represent zones of slickensides as much as 5 inches thick. Each such zone included two, and two zones included three, slickensided surfaces with different bearings. Five of these 14 bearings were in the N. 70°–93° W. arc, 6 between N. 30° W. and

different times. During a single compressive phase, movement in one direction might have met resistance, with deflection in a direction of less resistance. If this were true, one would expect these deflections to occur in a more random distribution. Another possible explanation is that stress directions changed. The northwestern Vermont region has been compressed during the Taconian, Acadian, and possibly the Appalachian orogenies. Presumably their stress directions were not identical,

yet each could have been relatively constant over the area. Renewed stress in a different direction would probably induce movement upon many old surfaces of weakness, and older slickensides would be obliterated, replaced by those oriented in a newer direction unless the new shear occurred on a different surface within the same zone. In most shears, then, slickensides indicate only the most recent direction of movement. Direction of movement for older compressions is indicated only on those surfaces where there was no renewed movement.

There is a striking correlation between slickenside bearings in the shales of this area and those Balk measured (1927, p. 62) on cross and glide joints in the granitic intrusives of Bethel, Barre, and Woodbury, Vermont, about 55 miles to the east. Balk found that, despite the different strikes of the joints, 61 per cent of the 460 slickensides were between N. 40° W. and N. 60° W. (compared with 59 per cent of 119 bearings between N. 35° W. and N. 60° W. in the shales of this area). Overthrust movement toward the northwest was repeatedly measured on the joints (Balk, 1927, p. 62, 91). He states (1927, p. 95) that "unmistakable evidence was found of a tectonic pressure that was directed toward the west or northwest, and that the magma itself had been notably affected by it." Cady has summarized opinions (1945, p. 580) that these intrusives are generally considered to be Devonian.

Fracture cleavage.—Fracture cleavage is nearly everywhere present in the more argillaceous beds of the Stony Point and Iberville formations. The term fracture cleavage is used here only in a descriptive sense, as defined by Swanson (1941, p. 1247), "the structure is due to closely spaced planes of parting a certain small distance apart," and "as a rule it is possible to see that the rock between the planes of parting . . . has no structure parallel to them, or at most any parallel structure is confined to a thin film along the parting planes." In Stony Point and Iberville shales, cleavage planes are more closely spaced in belts of intense folding, and, under the same structural conditions, they are more closely spaced in the more argillaceous beds than in the more calcareous beds.

Fracture cleavage plates in the argillaceous limestone of the Stony Point formation commonly range from half an inch to 5 inches thick. Fracture cleavage in the calcareous shale is much finer, and in the noncalcareous shale of the Iberville the planes are so close that there is a resemblance to flow cleavage (Swanson, 1941, p. 1246). Yet in thin section cut perpendicular to the finest cleavage of the noncalcareous shale, the cleavage is composed of somewhat irregular and discontinuous jointlike fractures 0.02 to 0.06 mm apart. The minute argillaceous particles of the groundmass are randomly oriented, though the elongate, angular quartz fragments (0.007 mm ±), sericite (0.003 to 0.04 mm long), and carbonaceous flakes (up to 0.07 mm long) more commonly parallel the bedding. Displacements of 0.01 to 0.04 mm occur along the cleavage planes.

Normally the fracture cleavage fans outward from the axis in crests and troughs of folds. Coarse cleavage fans more widely than the fine cleavage. Different cleavage dips occur in alternating beds where thin-cleaved calcareous shale is interbedded with coarsely cleaved argillaceous limestone, though the lines of intersection between cleavage and bedding for both sets approximately parallel the axes of the folds on which they occur. Cleavage plates on many folds are rotated on the *b* axis of Sander (Fairbairn, 1949, p. 6) into attitudes more closely parallel to the axial planes, presumably by continued compression after the cleavage formed. This rotation is evidenced by steplike displacement of bedding traces on cleavage plates. Similar rotation close to minor thrusts (Pl. 3, fig. 3) suggests that the cleavage originated on folds which were later cut by thrusts.

Two intersecting sets of fracture cleavage are equally well developed in an argillaceous limestone bed of the Stony Point formation on the east shore of Grand Isle, 1.75 miles N. 8° W. from the northwest tip of Savage Island. Attitudes of the cleavages, N. 5° E.–75° E. and N. 65° E.–45° SE., suggest compressions directed N. 35° W. and N. 85° W. at different times, but no regional conclusions can be drawn from one such occurrence.

Stylolites occur along almost vertical coarse fracture cleavage in the argillaceous limestone on Coon Point, south Alburg. Inasmuch as the rock must have been indurated at the time of

the folding which caused the fracture cleavage, the stylolites formed in hard limestone, probably under later compression in the manner described by Blake and Roy (1949, p. 787).

Minor folds.—The shales are most intensely folded, crumpled, and sheared close below the major thrusts along the east shore of Lake Champlain. Minor shearing is so extensive that major folding, if present, is impossible to delineate. More competent limestone interbeds are commonly buckled or stretched and broken, and their lenticular fragments separated by the more mobile shale (Pl. 3, fig. 4). Minor folds generally plunge northeastward and are overturned toward the northwest; some plunge south to southwestward. Minor folds plunging east to southeast, roughly parallel to the line of thrust movement (*a* axis), are not uncommon though nowhere are they more than a mile or so from the major thrusts. Figure 4 of Plate 4 shows folds plunging 46° S. 48° E. in buff to light-gray weathering, dark-gray laminated limestone interbedded with grayish-black calcareous shale close below the Champlain thrust (*cf.* Balk 1936, p. 738; Cloos, E., 1946, p. 26–48). In the present case the *a* axis folding seems to be the result of converging movement of masses of rock of varying resistance to compression, bounded by minor shears, as they were squeezed and mashed by the overriding massive thrust slice.

Summary of Depositional History of Shales

The fossiliferous (brachiopods, bryozoans, trilobites) Isle la Motte formation and the Larrabee and Shoreham members of the Glens Falls formation represent the lowest part of the Trenton group in the field area (Kay, 1937, p. 262, 264). In western South Hero the Isle la Motte has 20 feet of massively bedded limestone; the Larrabee, 72 feet of thin-bedded, somewhat shaly limestone; and the Shoreham 30 feet of somewhat thicker-bedded limestone with shale partings.

Above the Shoreham and below the Stony Point, 145 feet of Cumberland Head formation in western Grand Isle is composed of intercalated limestone and calcareous shale, predominantly limestone in the lower 30 feet, with an increasing proportion of shale upward (Kay, 1937, p. 274). Fossils are scarce, but some trilobites, brachiopods, and graptolites have been found (Ruedemann, 1921b, p. 98). The Cumberland Head is characterized by wavy, irregular bedding with occasional lateral gradation from limestone beds into calcareous shale.

In the Stony Point formation the lower 208 feet is calcareous shale (71 per cent), with intercalated thin limestone beds (29 per cent). The next 400 feet is made up of laminated argillaceous limestone (66 per cent) interbedded with calcareous shale (29 per cent) and hard, purer fine-grained limestone (5 per cent) in a somewhat cyclic pattern. Current bedding indicates currents flowing northeastward. Above this predominantly argillaceous limestone zone, the proportion of calcareous shale increases. The 239 feet near the top of the Stony Point is composed entirely of calcareous shale. The Stony Point has a very sparse fauna of trilobites, graptolites, and a few brachiopods (Ruedemann, 1921a, p. 112). Its total thickness is estimated to be 1000–1500 feet.

In the lower Iberville, noncalcareous shale is intercalated with calcareous shale, and the occasional hard beds are dolomite. The proportion of calcareous shale decreases upward. At some unknown distance above the base of the Iberville, a section of at least 730 feet shows cyclic interbedding of shale and graded dolomitic siltstone; the siltstone commonly shows current cross-lamination. The currents flowed southward to westward. In the thicker graded beds there is intrastratal crumpling. The author found no fossils in the Iberville, but Clark recovered graptolites of Utica age from the Iberville portion of a well core taken 4 miles north of the international boundary (Clark and Strachan, 1955, p. 688–697).

Overlying the Iberville is the Hathaway formation, composed of argillite and bedded radiolarian chert, chaotically deformed, with included masses of limestone, dolomite, dolomitic quartz siltstone, quartz sandstone, coarse graywacke, and chert.

The predominantly shallower, clear-water deposition of the fossiliferous lower Trentonian limestones changed gradually to deposition in deeper, more turbid waters with an increase in terrigenous content of the sediments and decrease in lime, a change in environment increasingly hostile to life.

Turbidity currents can account for cyclic deposition in the Iberville, and possibly also in the Stony Point. Indications of such currents and minor intrastratal slumping in the Iberville suggest a sloping bottom, or a bottom in deep water near a broad slope (clinoform of Rich, 1951, p. 6–7). The change in current direction from Stony Point (northeastward flowing) to Iberville (southward to westward) suggests that shallower waters during the deposition of the Iberville lay northeast or east of this field area, and that this slope developed with the beginning of Iberville deposition.

Large-scale, chaotic deformation of the Hathaway, with evidence for penecontemporaneous sliding and crumpling, and the coarse clastic blocks which include grains of metamorphic or igneous rock indicate that the slope may have developed into a submarine fault scarp from which the sliding started. The matrix of siliceous argillite and the radiolarian chert in the Hathaway indicate a fairly high percentage of silica in the water in which they were deposited, either in the field area or in the area where the sliding started (presumably to the northeast or east).

It is significant that the younger sediments seem to have been deposited in progressively deeper water. Farther south in the Champlain-Hudson Valley, in the Albany-Schenectady area of New York, the Snake Hill and Schenectady formations of middle Trenton age total about 6000 feet of shale and sandstone which become increasingly coarse in the younger part of the section. Lithology of the upper Schenectady suggests deposition in near-shore, shallow-water environments which became graded prior to Utica time. Deposition ceased about the close of Canajoharie-Sherman Fall time (Kay, 1937, p. 273–274). Thus, the rising middle Ordovician land, Vermontia (Kay, 1937, p. 290), must have been much closer to the Albany-Schenectady area than to the northern Champlain Valley, or at least higher near the Albany-Schenectady area than near the northern Champlain Valley.

Presence of chert and graywacke in the Hathaway indicates that products from a very different sedimentary environment have been dumped into the miogeosynclinal Champlain trough (Kay, 1937, p. 290) by submarine slides. Vermontia, separating the Champlain from the eugeosynclinal Magog trough, may not have extended as far north as the Canadian border at the time of deposition of the Hathaway. Sliding may have started on a submarine ridge or scarp extending northward from Vermontia.

Summary of Deformation of Shales

The shales are complexly folded and sheared, with fold axes trending a little east of north in the southern part of the area, and swinging more toward the northeast (N. 20°–30° E.) in the north. Although elongate narrow belts of intense deformation parallel fold trends, separated by broader belts of more gentle folding, general intensity of deformation increases toward the Champlain and Highgate Springs thrusts. Underlying more competent limestones and dolomites are gently folded, and tilted along normal faults; they are exposed only in the western part of the area, where the shale deformation is also more gentle.

Bearings of slickensides on bedding surfaces and other low-angle shears are surprisingly constant regardless of attitude of the shears; less than 3 per cent of the bearings are outside a 70 degree arc (N. 25° W. to S. 85° W.), and 59 per cent of the bearings lie within a 25 degree arc (N. 35° W. to N. 60° W.). Each of 6 shear zones in the shales included 2 or 3 surfaces with different directions of slickensides. Bearings of slickensides on 14 surfaces were measured in these 6 zones: 5 are between N. 70° W. and N. 93° W., 6 between N. 30° W. and N. 40° W., and 3 are N. 55° W. Bearings of slickensides in the shales correlate strikingly with those Balk recorded (1927, p. 62) in granitic intrusives 55 miles to the east, across the Green Mountain axis, which are considered Devonian.

Faults are abundant but of minor displacement and are related to compression which deformed the shales. Most are low-angle shears whose slickensides are approximately perpendicular to the nearest fold axes. Normal faults typical of the Adirondack flanks cannot be recognized in the shales studied, though they do occur in underlying more competent limestone and dolomite exposed in southwestern South Hero. However, an east-west-trending normal fault of considerable displacement, downthrown on the north, cuts tightly folded Stony Point beds at Wings Point, in western Charlotte, Vermont, about 14 miles south of the map area.

A normal fault of similar displacement cuts a bostonite dike intruded into Stony Point shale exposed in Lewis Creek at North Ferrisburg, Vermont, 17 miles south of the map area. Bostonite dikes also intrude the large thrust slices east of the Champlain fault.

Interpretation of the sequence of tectonic events is difficult because of inadequate and conflicting data. Cady (1945, p. 579–581) has summarized and evaluated evidence bearing upon this problem, stating that the Taconic allochthone, at the north end of the Taconic Range about 50 miles south of the area studied, was emplaced between middle Ordovician and late Silurian time and that the foreland thrusts, including the Champlain and Highgate Springs, may be continuous southward with thrusts that cut both the Taconic allochthone and middle Devonian beds in the Hudson Valley region (Schuchert and Longwell, 1932, p. 324). He suggests they are genetically related to the granitic intrusions which are "generally considered to be of Devonian age" (Clark, 1934, p. 16; Cooke, 1937, p. 85). On the other hand, Kay (1951) describes Taconic slates in the north end of the Taconic Range which lie on a plate of overturned Beldens marble in the Sudbury nappe, which in turn bevels the core of the Middlebury synclinorium. Some folds of the Middlebury synclinorium are also truncated by the foreland thrusts (Cady, 1945, p. 575). This might indicate that the foreland thrusts are Taconic in age.

Within the area of study there is little evidence to help determine the age of the Champlain and Highgate Springs thrusts. Present structure of the shales is genetically related to these thrusts. This is especially true of slickensided shears. Correlation of slickensides with those in Devonian (?) granite east of the Green Mountains suggests a genetic relationship which could date the foreland thrusting as Devonian, but the shales and the granite are separated by the Green Mountain belt of intense deformation. The six shear zones, each of which include more than one slickenside direction, might indicate renewed compression from a different direction. Possibly these represent different orogenies. If so, initial deformation could have been Taconic, with an approximate east–west axis of compression. A later compression on an axis closer to northwest–southeast would have replaced most of the earlier slickensides. The later compression might be Devonian.

Adirondack normal faulting is more recent than the foreland thrusting (Cady, 1945, p. 581) and seldom cuts the thrust slices or the younger shales in this area probably because the movement of the more competent older formations is taken up by internal crumpling in the more mobile overlying shale (Megathlin, 1938, p. 106).

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Appendix

Section 1

Northwest Grand Isle Section, Stony Point formation: measured along 1.8 miles of shore extending northward from a point 0.6 mile north of Wilcox Bay, western Grand Isle, Vermont. Column 1 gives the thickness of each unit. Column 2 gives the cumulative thickness to the top of each unit from the bottom of the section.

Covered

	Thickness (feet) (1)	(2)
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1)	1.0	634.6
Shale, calcareous, dark gray (N3)	0.4	633.6
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1)	0.4	633.2
Shale, calcareous, dark gray (N3)	0.9	632.8
Limestone, argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1) to light olive gray (5Y6/1), platy beds (¼ to ¾ in.)	6.3	631.9
Limestone, very fine-grained, uniform, dark gray (N3), weathering light olive gray (5Y2/1)	0.5	625.6
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1) to light olive gray (5Y6/1), thin-bedded (¼ to ¾ in.), with occasional 4- to 5-inch beds; with about 15 per cent dark-gray (N3) calcareous shale partings. The limestone here shows ripple drift cross-lamination: wave length 12 inches, amplitude 1 inch, crest strike N. 50° W., current indicated toward N. 40° E.	12.2	625.1
Limestone, very fine-grained, uniform, dark gray (N3), weathering light olive gray (5Y6/1)	0.4	612.9
Limestone, argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1) to light olive gray (5Y6/1), platey beds (¼ to ¾ in.); with dark gray (N3) calcareous shale partings (about 25 per cent)	16.7	612.5
Limestone, very fine-grained, faintly laminated, dark gray (N3), weathering light olive gray (5Y6/1)	0.3	595.8
Limestone, argillaceous, laminated, with shale partings, as in second unit above	31.2	595.5
Bedding-shear zone, finely brecciated and slickensided	0.5	564.3
Limestone, argillaceous, laminated, with shale partings, as in 16.7-foot unit, above	12.0	563.8
Shale, calcareous, dark gray (N3)	3.9	551.8
Limestone, silty and argillaceous, laminated with very fine cross-bedding, dark gray (N3), weathering olive gray (5Y4/1), platy beds (¼ to 4 in.)	15.4	547.9
Bedding-shear zone, finely brecciated and slickensided	0.4	532.5
Limestone, fine-grained crystalline, medium gray (N5), weathering medium light gray (N6), single bed of varying thickness (½ to 1½ in.)	0.1	532.1
Limestone, silty and argillaceous, laminated, with fine cross-bedding, dark gray (N3), weathering olive gray (5Y4/1) to light olive gray (5Y6/1) (4 to 10 in.; about 70 per cent); with dark-gray (N3) calcareous shale partings (about 30 per cent). When the limestone beds reach 8 to 10 inches, there is in the middle of the bed a layer (about 1 in.) of very fine-grained purer limestone, dark gray (N3), weathering light olive gray (5Y6/1)	7.6	532.0
Limestone, argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), platy beds (¼ to 1 in.); with two interbedded very fine-grained purer limestone beds, dark gray (N3), weathering light olive gray (5Y6/1), each 3 to 4 inches	4.8	524.4
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), (5 to 9 in.; about 60 per cent); with dark-gray (N3) calcareous shale partings (about 40 per cent). Toward the upper part of this zone, the limestone beds become somewhat laminated, silty, and argillaceous	13.7	519.6
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), platy beds (¼ to 1 in.)	6.4	505.9
Limestone, very fine-grained, uniform, dark gray (N3), weathering light olive gray (5Y6/1)	1.0	499.5
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), platy beds (½ to 4 in.; about 60 per cent); with dark-gray (N3) calcareous shale partings (about 40%)	17.8	498.5
Limestone, very fine-grained, faintly laminated, dark gray (N3), weathering olive gray (5Y4/1) to light olive gray (5Y6/1); with lenticular pyrite concretions (up to ¼ by ¾ in.) common through the mid-portion	0.1	480.7

	Thickness (feet) (1)	(2)
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), platy beds (¼ to 3 in.; about 85 per cent); with dark-gray (N3) calcareous shale partings (about 15 per cent)	22.6	480.6
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1) to light olive gray (5Y6/1), platy beds (¼ to 4 in.; about 50 per cent); interbedded with purer limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), in beds up to 9 inches, which occasionally are intricately cross-laminated on a very small scale	8.5	458.0
Shale, calcareous, dark gray (N3) (about 70 per cent); interbedded with laminated silty and argillaceous limestone, dark gray (N3), weathering olive gray (5Y4/1), (¼ to 1½ in.; about 30 per cent)	13.7	449.5
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), (¼ to 2 in.)	1.5	435.8
Limestone, very fine-grained, cross-laminated, dark gray (N3), weathering light olive gray (5Y6/1). In this bed there is ripple drift cross-lamination: wave length 18 inches, amplitude 1½ inches, crest strike N. 70° W., current indicated toward N. 20° E.	0.4	434.3
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), platy beds (¼ to 2 in.)	6.9	433.9
Limestone, very fine-grained, faintly laminated (on weathered surfaces), dark gray (N3), weathering olive gray (5Y4/1) to light olive gray (5Y6/1), (6 to 8 in.; about 60 per cent); with intercalated dark-gray (N3) calcareous shale (4 to 5 in.; about 40 per cent)	12.4	427.0
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), (½ to 2 in.; about 65 per cent); interbedded with purer limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), (¾ to 2 in.; about 20 per cent); with partings of dark-gray (N3) calcareous shale (about 15 per cent)	10.3	414.6
Covered	16.5	404.3
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), platy beds (¼ to ¾ in.; about 85 per cent); interbedded with purer limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), (½ to 3 in.; about 15 per cent)	11.0	387.8
Covered	9.5	376.8
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), two 6-inch beds; with interbedded dark-gray (N3) calcareous shale	1.5	367.3
Covered	4.5	365.8
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), platy beds (¼ to ¾ in.)	1.0	361.3
Covered	5.5	360.3
Limestone, silty and argillaceous, as in 1.0-foot unit second above	16.6	354.8
Bedding shear zone, finely brecciated and slickensided	1.0	338.2
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), (¼ to 5 in.; about 75 per cent); with partings of dark-gray (N3) calcareous shale (about 25 per cent)	28.5	337.2
Shale, calcareous, dark gray (N3) (about 80 per cent); with limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), platy beds (¼ to ½ in.; about 20 per cent)	20.1	308.7
Shale, calcareous, dark gray (N3) (about 50 per cent); with limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), (¼ to 5 in.; about 50 per cent)	5.5	288.6
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1)	0.5	283.1
Limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), (¼ to 3 in.; about 60 per cent); with intercalated dark-gray (N3) calcareous shale (about 40 per cent)	2.6	282.6
Shale, calcareous, dark gray (N3; about 60 per cent); with intercalated limestone, silty and argillaceous, laminated, dark gray (N3), weathering olive gray (5Y4/1), (¼ to 1 in.; about 40 per cent)	7.1	280.0

	Thickness (feet) (1)	(2)
Covered	16.5	272.9
Shale, calcareous, dark gray (N3)	6.0	256.4
Shale, calcareous, dark gray (N3; about 60 per cent); with intercalated limestone, silty and argillaceous laminated, dark gray (N3), weathering olive gray (5Y4/1), (½ to 2 in.; about 40 per cent)	14.9	250.4
Covered	37.4	235.5
Shale, calcareous, dark gray (N3; about 90 per cent); with limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), commonly in 1 to 4 inch beds, but occasionally as thick as 10 inches, occurring every 4 feet or so (about 10 per cent)	49.6	198.1
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), in beds (6 to 8 in.; about 80 per cent); with dark-gray (N3) calcareous shale partings (about 20 per cent)	3.9	148.5
Shale, calcareous, dark gray (N3; about 80 per cent); with intercalated limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), (1 to 3 in.; about 20 per cent)	12.4	144.6
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), in beds (6 to 18 in.; about 85%); with occasional dark-gray (N3) calcareous shale partings (2 to 4 in.; about 15 per cent)	7.0	132.2
Shale, calcareous, and limestone, very fine-grained, as in 12.4-foot unit, second above	27.2	125.2
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), in beds of 6 to 10 inches	4.4	98.0
Shale, calcareous, dark gray (N3)	21.9	93.6
Shale, calcareous, dark gray (N3) (about 80 per cent); with intercalated limestone, very fine grain, dark gray (N3), weathering light olive gray (5Y6/1), (¾ to 2 in.; about 20 per cent)	4.4	71.7
Shale, calcareous, dark gray (N3)	4.0	67.3
Covered	18.4	63.3
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1)	0.2	44.9
Shale, calcareous, dark gray (N3)	11.2	44.7
Covered	3.0	33.5
Shale, calcareous, dark gray (N3; about 95 per cent); with intercalated limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), (½ to ¾ in.; about 5 per cent)	8.0	30.5
Shale, calcareous, dark gray (N3)	3.8	22.5
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), in two beds (about 55 per cent); with intercalated calcareous shale (about 45 per cent)	1.3	18.7
Shale, calcareous, dark gray (N3)	2.6	17.4
Limestone, very fine-grained, dark gray (N3), weathering light olive gray (5Y6/1), (3 to 7 in.; about 70 per cent); with intercalated calcareous shale (about 30 per cent)	4.2	14.8
Shale, calcareous, dark gray (N3)	1.0	10.6
Limestone, fine-grained, dark gray (N3), bearing fragments of trilobites, brachiopods, and orthocone cephalopods	0.3	9.6
Shale, caclareous, dark gray (N3)	9.3	9.3

Cumberland Head formation.

Section 2

South Alburg Section, Stony Point formation: measured on the shore line of the promontory west of Coon Point, south Alburg, Vermont. This section correlates well both lithologically and faunally with Ruedemann's Stony Point "type" section (1921a, p. 112) and resembles the upper part of the northwest Grand Isle section, probably including younger beds than are exposed there. Column 1 gives the thickness of each unit. Column 2 gives the cumulative thickness to the top of each unit.

Covered

	Thickness (feet) (1)	(2)
Limestone, argillaceous and occasionally silty, dark gray (N3), weathering olive gray (5Y4/1) to greenish gray (5GY6/1) or light olive gray (5Y6/1) in darker and lighter bands and laminae ($\frac{1}{16}$ to 3 in.)	40.0+	258.74+
Limestone, fine-grained, crystalline, medium gray (N5), weathering medium light gray (N6)	0.07	218.74
Limestone, argillaceous and silty, as in 40.0-foot unit at top of section	0.5	218.67
Limestone, very fine-grained, uniform, dark gray (N3), weathering light olive gray (5Y6/1)	0.1	218.17
Limestone, argillaceous and silty, as in 40.0-foot unit at top of section	1.1	218.07
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.2	216.97
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	4.8	216.77
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.1	211.97
Limestone, argillaceous and silty, as in 40.0-foot unit at section top. *Triarthrus becki* common	11.4	211.87
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.4	200.47
Limestone, argillaceous and silty, as in 40.0-foot unit at section top. *Triarthrus becki* fairly common in upper part	6.9	200.07
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.1	193.17
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	22.5	193.07
Limestone, fine-grained, crystalline, as in 0.07-foot unit next to section top	0.05	170.57
Limestone, argillaceous and silty, as in 40.0-foot unit at section top. *Triarthrus becki* very common in upper part	5.7	170.52
Limestone, very fine-grained, uniform, dark gray (N3), weathering light olive gray (5Y6/1), with small pyrite concretions scattered along central portion of bed	0.3	164.82
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	20.0	164.52
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.8	144.52
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	2.0	143.72
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.4	141.72
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	0.8	141.32
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.4	140.52
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	0.5	140.12
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.5	139.62
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	2.7	139.12
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.1	136.42
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	0.4	136.32
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.4	135.92
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	18.2	135.52
Limestone, argillaceous and silty, as in 40.0-foot unit at section top. *Triarthrus becki* very common	4.0	117.32
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	12.6	113.32
Limestone, very fine-grained, uniform, as in 0.1-foot unit near section top	0.7	103.72
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	2.5	100.02
Limestone, very fine-grained, uniform, dark gray (N3), weathering light olive gray (5Y6/1), ($\frac{3}{4}$ to 2 in. beds; about 80 per cent); with intercalated 1-inch beds of argillaceous and silty limestone	1.0	97.52
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	1.8	96.52
Limestone, fine-grained, uniform, as in 0.1-foot unit near section top	0.3	94.72
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	6.1	94.42
Covered	6.0	88.32
Limestone, fine-grained, crystalline, as in 0.07-foot unit next to section top	0.02	82.32

	Thickness (feet) (1)	(2)
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	3.7	82.30
Limestone, fine-grained, uniform, as in 0.1-foot unit near section top	0.1	78.60
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	5.2	78.50
Limestone, very fine-grained, uniform, dark gray (N3), weathering light olive gray (5Y6/1), with small pyrite concretions scattered through mid-portion	0.2	73.30
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	1.8	73.10
Limestone, very fine-grained, uniform, with pyrite concretions, as in unit second above	0.3	71.30
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	26.0	71.00
Limestone, fine-grained, crystalline, as in 0.07-foot unit next to section top	0.1	45.00
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	0.3	44.90
Limestone, very fine-grained, uniform, dark gray (N3), weathering light olive gray (5Y6/1), with small pyrite concretions scattered through the mid-portion of the bed	0.1	44.60
Limestone, argillaceous and silty, as in 40.0-foot unit at section top	11.5	44.50
Limestone, very fine-grained, uniform, with pyrite concretions, as in bed second above	0.3	33.00
Limestone, argillaceous and silty, as in 40.0-foot unit at section top, with a few *Triarthrus becki*	9.7	32.70
Limestone, argillaceous and silty, as in 40.0-foot unit at section top, with occasional horizons along which there are scattered small pyrite concretions	23.0	23.00
Covered		

Section 3

Clark Point Section, Iberville formation: measured along the western shore of Clark Point, southern Hog "Island," in western Swanton, Vermont. Column 1 gives the thickness of each unit. Column 2 gives the cumulative thickness to the top of each unit.

Covered		
Shale, thin cleaving, dark gray (N3), in very regular beds (1 to 2 in.), with thin (⅛ to ¾ in.) grayish-black (N2) bands in the upper part of each bed; usually interbedded with dolomitic siltstone, grading finer upward through a series of laminae, medium dark gray (N4), weathering moderate yellowish brown (10YR5/4) to olive gray (5Y3/2), (beds ⅛ to ¼ in.)	38.5	303.7
Shale, silty, thin cleaving, dark gray (N3), (¾ to 2 in. beds, about 60 per cent); interbedded with dolomitic siltstone, laminated, medium dark gray (N4), weathering moderate yellowish brown (10YR5/4) to olive gray (5Y3/2) (1 to 1½ in. beds, about 40 per cent)	2.3	265.2
Shale, thin cleaving, dark gray (N3), (1 to 2 in. beds), with occasional very thin (⅛ to ¼ in.) dolomitic siltstone beds, which grade finer upward through a series of laminae, medium dark gray (N4), weathering moderate yellowish brown (10YR5/4) to olive gray (5Y3/2)	45.8	262.9
Shale, thin cleaving, dark gray (N3) (about 60 per cent); with interbedded laminated silty dolomite, medium dark gray (N4), weathering moderate yellowish brown (10YR5/4), which grades upward into the shale (about 40 per cent), and also grades laterally into shale within 50 feet	0.7	217.1
Shale, thin cleaving, dark gray (N3), in very regular 1½ to 2½-inch beds (occasionally as thick as 4 in.); which are separated by graded and laminated dolomitic siltstone, medium dark gray (N4), weathering moderate yellowish brown (10YR5/4) to olive gray (5Y3/2), (beds ¼ to ⅝ in.). Through occasional zones of 2 to 3 feet, there is no dolomitic siltstone	55.7	216.4
Dolomite, silty, banded and laminated, medium dark gray (N4), weathering moderate yellowish brown (10YR5/4) to dark yellowish brown (10YR4/2), with a shale parting in the middle (¼ in.)	1.0	160.7
Shale, thin cleaving, and dolomitic siltstone, as in 55.7-foot unit above	18.0	159.7
Siltstone, dolomitic, laminated and banded, medium dark gray (N4), weathering dark yellowish brown (10YR4/2), (1 to 5 in. beds, about 50 per cent); with intercalated shale, thin cleaving, dark gray (N3), (¾ to 2 inch beds, about 50 per cent)	2.4	141.7

	Thickness (feet) (1)	(2)
Shale, thin cleaving, and dolomitic siltstone, as in 55.7-foot unit above	74.5	139.3
Dolomite, fine-grained, uniform, medium gray (N5), weathering moderate yellowish brown (10YR5/4), banded with laminated silty dolomite which shows ripple drift cross-lamination; and argillaceous dolomite, dark gray (N3), weathering dark yellowish brown (10YR4/2). Southward to westward current indicated	1.1	64.8
Shale, thin cleaving, and dolomitic siltstone, as in 55.7-foot unit above	22.5	63.7
Siltstone, dolomitic, ripple drift cross-laminated, medium dark gray (N4), weathering moderate yellowish brown (10YR5/4) to dark yellowish brown (10YR4/2), grading upward into fine-grained uniform dolomite. Southward to westward currents indicated by ripple drift	0.2	41.2
Shale, thin cleaving, and dolomitic siltstone, as in 55.7-foot unit above	2.8	41.0
Dolomite, very fine-grained, uniform, medium gray (N5), weathering moderate yellowish brown (10YR5/4)	0.2	38.2
Shale, thin cleaving, and dolomitic siltstone, as in 55.7-foot unit above	9.8	38.0
Dolomite, very fine-grained, uniform, medium gray (N5), weathering moderate yellowish brown (10YR5/4)	0.2	28.2
Shale, thin cleaving, and dolomitic siltstone, as in 55.7-foot unit above; except that in some of the thicker siltstone beds there is ripple drift cross-lamination. Southward to westward flowing current is indicated	21.2	28.0
Dolomite, very fine-grained, uniform, dark gray (N5), weathering moderate yellowish brown (10YR5/4)	0.6	6.8
Shale, thin cleaving, and dolomitic siltstone, as in 55.7-foot unit above	2.5	6.2
Dolomite, very fine-grained, uniform, medium gray (N5), weathering moderate yellowish brown (10YR5/4), with two 1-inch dark yellowish-brown (10YR4/2) weathering laminated silty bands	0.7	3.7
Shale, thin cleaving, and dolomitic siltstone, as in 55.7-foot unit above	3.0	3.0
Covered		

Section 4

Northern Woods Island Section, Iberville formation: measured along the north-central portion of the western shore of Woods Island, in northeastern Lake Champlain, in St. Albans Township.

The oldest beds of this section have the same attitude as the youngest beds of the southern Woods Island section and lie along their strike but are separated from them by 1000 feet of beach cover. There is no suggestion of dislocation in the covered area; if none exists, there should be an approximate correlation between the top of the southern section and the bottom of the northern.

In both Woods Island sections there is probably some minor thickening by solid flow in the shales, as evidenced by occasional compressional rupture, slip, and minor overlap in the dolomite beds. While it is recognized that these sections may be somewhat thicker than the true predeformational thickness, they are the only measurable exposures available. Column 1 gives the thickness of each unit. Column 2 gives the cumulative thickness to the top of each unit.

Covered		
Shale, thin cleaving, with fine dark-gray (N3) and gritty grayish-black (N2) bands (½ to 2 in.), in 1- to 6- inch beds (about 96 per cent); with frequent (every 4 to 8 in.) laminated dolomitic siltstone beds (⅛ to ½ in.), less frequently (every 2 to 3 ft.), as thick as 3 inches, medium dark gray (N4), weathering moderate yellowish brown (10YR5/4) to olive gray (5Y3/2) (about 4 per cent). The dolomitic siltstone beds occasionally show current cross-lamination	19.0	363.2
Dolomite, very fine-grained, uniform, medium gray (N5), weathering moderate yellowish brown (10YR5/4), single bed	0.3	344.2
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	20.7	343.9
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.2	323.2
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	31.6	323.0
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.2	291.4
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	0.9	291.2

	Thickness (feet) (1)	(2)
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.3	290.3
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	13.0	290.0
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.3	277.0
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	3.9	276.7
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.2	272.8
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	22.2	272.6
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.3	250.4
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	16.9	250.1
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.1	233.2
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	6.0	233.1
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.6	227.1
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	1.8	226.5
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.2	224.7
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	8.2	224.5
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.2	216.3
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	27.1	216.1
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.4	189.0
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	18.1	188.6
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.5	170.5
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	46.9	170.0
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.3	123.1
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	12.2	122.8
Dolomite, argillaceous, fracture cleaved, medium dark gray (N4), weathering dark yellowish brown (10YR4/2), single bed	0.5	110.6
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	30.0	110.1
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.6	80.1
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	1.8	79.5
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.7	77.7
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above	31.9	77.0
Dolomite, very fine-grained, uniform, as in 0.3-foot unit above	0.6	45.1
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 19.0-foot unit above. Three of the thicker (3 in. ±) dolomitic siltstone beds in this unit show ripple drift cross-lamination, indicating a southward current flow	44.5	44.5
Covered		

Southern Woods Island Section, Iberville formation: measured along the south-central portion of the western shore line of Woods Island. The youngest beds of this section are presumed to be roughly correlative with the oldest beds of the northern Woods Island section.

	Thickness (feet) (1)	(2)
Covered		
Shales, thin cleaving, with fine dark-gray (N3) and gritty grayish-black (N2) bands (¼ to 2 in.), in 1- to 5-inch beds (about 96 per cent); with frequent (every 8 to 24 in.) beds (⅛ to 2 in.) of dolomitic siltstone, laminated, medium dark gray (N4), weathering moderate yellowish brown (10YR5/4) to olive gray (5Y3/2) (about 4 per cent). The dolomitic siltstone beds occasionally show current cross-lamination	14.1	369.1
Dolomite, very fine-grained, uniform, medium gray (N5), weathering moderate yellowish brown (10YR5/4), single bed	0.5	355.0
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	61.5	354.5
Dolomite, very fine-grained, medium gray (N5), weathering moderate yellowish brown (10YR5/4), grading laterally into shale within 25 feet	0.2	293.0
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	3.0	292.8
Dolomite, very fine-grained, uniform, as in 0.5-foot unit above	0.7	289.8
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	49.6	289.1
Dolomite, very fine-grained, uniform, as in 0.5-foot unit above	0.7	239.5
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	38.2	238.8
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above, except some of the shale beds are as thick as 8 inches	12.0	200.6
Dolomite, very fine-grained, uniform, medium gray (N5), weathering moderate yellowish brown (10YR5/4), in the lower part, grading upward into argillaceous dolomite with fracture cleavage, medium dark gray (N4), weathering dark yellowish brown (10YR4/2)	0.5	188.6
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	6.5	188.1
Dolomite, fine-grained, uniform, grading upward into argillaceous dolomite, as in second unit above	0.4	181.6
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	46.5	181.2
Dolomite, very fine-grained, uniform, as in 0.5-foot unit near section top	0.5	134.7
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	43.5	134.2
Dolomite, very fine-grained, uniform, as in 0.5-foot unit near section top	0.9	90.7
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	25.0	89.8
Dolomite, very fine-grained, uniform, as in 0.5-foot unit near section top	1.0	64.8
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	13.0	63.8
Dolomite, very fine-grained, uniform, as in 0.5-foot unit near section top	0.3	50.8
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above	18.7	50.5
Dolomite, very fine-grained, uniform, as in 0.5-foot unit near section top	0.8	31.8
Shale, thin cleaving, banded, with dolomitic siltstone interbeds, as in 14.1-foot unit above. Ripple drift cross-lamination was found in one 3-inch dolomitic siltstone bed, indicting a southward flowing current	31.0	31.0
Covered		

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ORIGIN OF CARBONATE CONCRETIONS IN SHALES, MAGDALENA VALLEY, COLOMBIA

By L. G. Weeks

Abstract

Carbonate concretions that include marine organisms such as ammonites or even the full body of a fish are abundant in Cretaceous shales of the Magdalena Valley of Colombia, South America. The occurrence of carbonate concretions in central basin shale facies is common not only throughout the Cretaceous of the lengthy Andean geosyncline but also in similar facies throughout the world. The concretions are of syngenetic or early diagenetic origin. They formed soon after deposition, when the now compact muds were plastic enough to flow in around them. The fish and other organisms calcified rapidly, and the enveloping concretions developed sufficiently early to arrest extreme decomposition and to withstand compression from the many thousands of pounds of subsequent overburden.

The central-basin environments in which most carbonate concretion-bearing shales were deposited were not favorable to carbonate deposition. Such environments are stagnant and so are charged with carbon dioxide while the pH is too low for calcium carbonate deposition. However, an alkalinity adequate for such deposition in calcium-ion-rich waters may be locally created by the ammonia that evolves rapidly with decomposition of proteinaceous (nitrogen-bearing) organic matter.

Similarly, limited decomposition, mainly anaerobic, of organic matter that impregnates deeper basinal muds may create enough alkalinity to free calcium carbonate from the lime-bearing solutions permeating the muds. This may account for calcareous shales in basins where pure limestones are restricted to the aerated shallow flank, shelf, or lesser bottom-high areas.

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Introduction

Fossil fish have been discovered in carbonate concretions which occur in great abundance in Cretaceous shales in the middle Magdalena Valley of Colombia, South America (Fig. 1). The specimen figured here (Pl. 1, figs. 1 and 2) was collected in Quebrada Peralonzo in the Department of Magdalena about 17 kilometers S.40°E. from Aguachica on the Gamarra-Ocana road.

Acknowledgments

Mr. Donald J. Podesta, geologist for International Petroleum (Colombia) Limited, dis-

covered the fossil fish and provided photographs of typical concretions and shales of the discovery locality. Mr. Podesta and Mr. E. D. Ackerman, Exploration Department Manager for Intercol, described the fossil-fish occurrence.

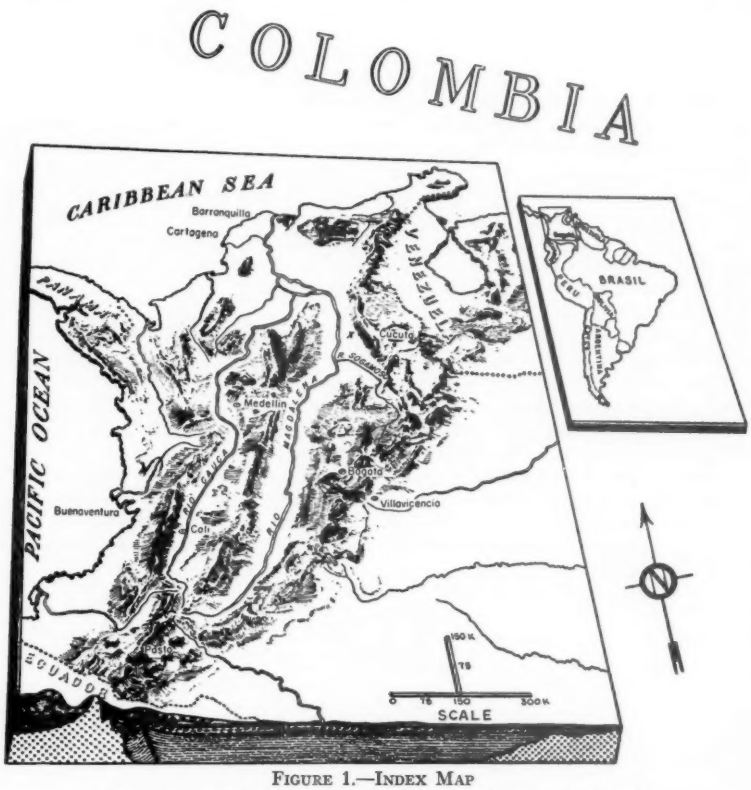

FIGURE 1.—INDEX MAP
Showing locality (x) where fish-bearing concretions were collected

The writer is indebted to both these gentlemen and to International Petroleum (Colombia) Limited for their interest in having this account published and for permission to deposit the specimen with the collection of fossil fishes in the American Museum of Natural History, New York. Mr. William D. Meyers, International Petroleum (Colombia) Limited, supplied the photographs for Figure 2 of Plate 2 and Figure 1 of Plate 4; Mr. M. Dean Williams, of the same company, presented the photograph for Figure 2 of Plate 3. Thanks are due the Public Relations Department of the Standard Oil Company (N. J.) for making and supplying the photographs of the fish for this paper. Dr. Bobb Schaeffer, Associate Curator of Fossil Fishes at the American Museum of Natural History, aided greatly in study of the fish specimen and in showing and discussing specimens from other areas, and he kindly permitted the writer to photograph the fish specimen from the Museum's collection (Pl. 4, fig. 2). The writer is grateful to Dr. Allen C. Tester and Dr. Heintz A. Lowenstam for critical reading of the text and for helpful suggestions, and to Mr. George Langford for information on the incidence of recognizable organic remains in concretions of the Pennsylvanian of Illinois. Professor Walter H. Bucher referred the writer to the experimental observations on animal carcasses on sea bottoms or burial at shallow depth in sediments

recorded by Franz Hecht (1933, p. 165–249).

Environment and Lithologic Relationships

The concretion-bearing shales of Colombia are part of a thick Cretaceous marine series which includes limestones and well-defined zones of dark bituminous cherts. Some sand has developed mainly along the basin flanks, particularly at or near the base of the Cretaceous transgression, and in the late regressive phases of the Cretaceous cycle.

This great series of sediments accumulated in a geosyncline which extended along the present Andean belt from the island of Trinidad westward through Venezuela, and south and southwest through Colombia, Ecuador, and Peru. The thickest Cretaceous accumulations are now largely incorporated into the mountain rocks of the Tertiary-Quaternary Andean chain.

The shales of the fossil locality are middle Cretaceous, approximately early Albian. They are thinly plated, black, and calcareous. Similar dark, bituminous shales are found at many places along the Andean geosyncline, as are shaly limestones. Pure limestones are confined to areas of little terrigenous sediment influx, and are restricted to aerated bottom areas such as broad, shallow basin shelves; they may accumulate over more of the basin floor during periods of limited stagnation.

Description of Concretions

Concretions are abundant and range in diameter from a few centimeters to approximately 8 meters. They are generally spherical to oval or oblong and slightly flattened parallel to the bedding with which they are stratigraphically aligned in narrow, recurrent zones (Pl. 2, fig. 1). This alignment of concretions along a plane may indicate a diastem or brief period of nondeposition. Nodular growths of various compositions are known to have developed over areas of little or no deposition on present sea bottoms, and they not uncommonly mark unconformities or disconformities.

Most of the concretions are composed of carbonate, largely calcium carbonate, although some are argillaceous or siliceous. They are gray to black, usually dense and hard with a distinct petroliferous odor when freshly broken. Ammonites are their most common inclusion; these contain tarry, green to black oil droplets, oil blebs, or oil staining.

Carbonate concretions are generally associated over the world with a stagnant water shale, marl, or argillaceous limestone facies. The manner in which the shale bends around the concretions (Pl. 3, figs. 1 and 2; Pl. 4, fig. 1) is typical of the relationships throughout this area and indicates that such concretions developed soon after deposition.

With regard to this important point, Allen C. Tester (Personal communication) stated: "The reviewer has examined thousands of such concretions in many stratigraphic and geographic positions (including the Cretaceous sediments of Colombia) and has not seen any that show a settling or compression of the shale under the concretion. . . . This has been interpreted to mean that the mud was compacted sufficiently to bear the weight of the concretion as it grew. The curvature at the sides and over the top was accomplished by contemporaneous or penecontemporaneous growth of the concretion and of mud deposition."

It is true that Figure 1 of Plate 3 appears to indicate little, if any, compaction of the shale beneath the concretion. However, Figure 2 of Plate 3 and Figure 1 of Plate 4 show apparent compaction below as well as above the concretion.

In answer to a query in this regard, E. D. Ackerman replied that it is the observation of all geologists in his department that "bending above and below the concretions appears to be present in almost every case. However, in isolated instances incipient bedding planes appear to pass through the concretion with no deformation." The writer has studied concretions in many sections over the world, and it is his observation that deformation as a result of compaction is usually more pronounced above concretions than it is below them.

Description of Fossil Fish

Bobb Schaeffer (Personal communication) reported: "The specimen belongs to a family of teleost fishes called the Elopidae which includes the living ten-pounders, and, in some classifications, the tarpon. . . . Unfortunately, the speci-

men is not complete enough to make a positive generic identification. . . .

"The elopids from the late Cretaceous on the northeast coast of Brazil were considered by Agassiz in 1841. Perhaps the most important contributions are those of Jordan (1921,) Woodward (1901), and Dunkle (1940). The richest locality for these forms is at Barra do Jardim State of Ceara in the Formacao Santana of the Serie Araripe. Here the fishes occur, often 'in the round,'[1] in limestone concretions.

". . . Although the resemblance of this form to the Brazilian elopids suggests that it is Cretaceous in age, it is important to note that this family of fishes has had a conservative evolution and that it occurs widely in marine deposits from the early Cretaceous to the Quaternary."[2]

Other Fish-Bearing Concretions

Other fish-bearing concretions were described by Bobb Schaeffer: "There are numerous other localities, of various ages, that have fishes in concretions. For instance, the Upper Devonian Cleveland shale of Ohio has large limestone concretions that frequently contain arthrodire fishes. These are usually partially or completely flattened. This same formation also has fish bones embedded directly in the shale. Molds of uncrushed or partially compressed fishes have been found in the Middle Pennsylvanian beds at Mazon Creek, Illinois, in the Texas Permian, and in the Lower Triassic of Madagascar. The Upper Jurassic of western Cuba contains beautifully preserved fishes in black limestone concretions, and these require acid preparation or sectioning for study. The most recent examples of concretionary formation around fish remains are probably found in Greenland and northern Canada. These concretions are generally considered to be post-Pleistocene in age and are frequently shaped like the fish *Mallotus* contained therein" (*See* Pl. 4, fig. 2).

The fish undergo varying degrees of compression in concretion occurrences. However, any flattening or reduction in size probably resulted from partial decomposition prior to calcification rather than from compression. In the Madagascar Lower Triassic the fish are encased in marine calcareous sand concretions. In most of these specimens the body of the fish has disintegrated, apparently after formation of the concretion, leaving a hollow limonite-faced mold in the original form of the fish.

In the Upper Devonian Cleveland shales, the fish are found in black carbonate concretions, and bones of the fish are dispersed in the shales. This may indicate that a concretion-forming environment was not prevalent generally and, where existent, decay had partially been completed before it was halted by calcification and carbonate encasement. The fish-bearing black carbonate concretions of the Upper Jurassic of Cuba have formed in bituminous shales and shaly limestones; the environment of deposition was similar to that of the Colombia concretion-bearing sediments. Hundreds of examples of carbonate concretion-bearing shales, limy shales, or shaly limestones of similar deposition

[1] A phrase used to indicate that the specimen is full-bodied, that is, undiminished or very little diminished in form and dimensions from its original condition in life.

[2] Other fossils established the age of the shales in which the fish-bearing concretions occur as middle Cretaceous (Albian).

Plate 1.—FOSSIL FISH

Figure 1.—Embedded in carbonate concretion

Figure 2.—Removed from concretion. Note absence of flattening of fish body, indicating early lithification and formation of the enveloping carbonate concretion before decomposition or compression of the carcass could occur.

Plate 2.—CARBONATE CONCRETIONS

Figure 1.—Typical outcrop of dark-gray and brownish-gray carbonate concretions in dark-gray Cretaceous shales. This exposure is in the foothills near Las Bocas and the Rio Sogomosa on the east side of the Magdalena Valley, Santander, Colombia. Note the alignment of the concretions. Courtesy Allen C. Tester, State University of Iowa.

Figure 2.—Outcrop of Cretaceous La Luna formation showing several concretions. Courtesy of William D. Meyers, International Petroleum (Colombia) Limited.

FIGURE 1

FIGURE 2

FOSSIL FISH

FIGURE 1

FIGURE 2

CARBONATE CONCRETIONS

FIGURE 1

FIGURE 2

CARBONATE CONCRETIONS

FIGURE 1

FIGURE 2

CARBONATE CONCRETIONS

environment might be cited over the world. Similar concretions are found in the Cretaceous geosyncline of Mexico and in the Mowry, Carlile, and other Cretaceous shales of the Rocky Mountain and adjacent states. "Many of the carbonate concretions of the Carlile and other Upper Cretaceous formations are not pure carbonate but some are siderite" (A. C. Tester, personal communication). Siderite is a more common constituent of sediments and of nodules in them than generally believed, according to recent mineralogic studies of Paul Kerr of Columbia University (Personal communication).

Reviewers of this paper asked about the incidence of recognizable fossil matter in the concretions. The following quotation is from a letter by E. D. Ackerman, to whom this question was referred: "Estimates here of the percentage of concretions in the La Luna formation that have no recognizable fossils today range from 75–25 per cent. A likely average would be 40–50 per cent. The proportion that carry ammonites is somewhere in the neighborhood of 10–15 per cent. It should be pointed out, however, that in some zones almost every concretion carries recognizable fossils (up to 90 per cent) and in other zones the concretions are practically barren (less than 5 per cent)."

George Langford, of the Chicago Natural History Museum, replied with regard to the iron carbonate concretions in the Pennsylvanian series of Illinois: "I and those with me have split over one million nodules in the strip coal mines west of Wilmington, Will County, Illinois. Whether the nodules contain fossils or not depends upon collecting spots and upon observations. Ones which at first appear to be blanks will often be found to contain a fossil nucleus if re-split to destruction. At one of several small collecting areas, the fossil-bearing nodules will run as high as 90 per cent. In several other areas, hardly 10 per cent. . . . If there is any dependable rule, I don't know what it is. The proportions in all of the nodules I have split would probably run near 75 per cent fossiliferous, because I learned to avoid spots where there were many blanks."

Time of Concretion Development

Views on the time of origin of concretions may be grouped under two headings: (1) Syngenetic—formation at the time of deposition of the containing sediments (or very shortly thereafter—diagenetic), (2) Epigenetic—formation subsequent to the lithification of the containing strata.

Perhaps concretions have originated in both of these ways. An epigenetic origin has been most widely accepted, and it does explain why, in certain concretions in permeable strata, the bedding continues on through the concretion.

Where carbonate concretions are widespread and abundant in shales, calcareous shales, and argillaceous limestones, a syngenetic or early diagenetic origin is more probable, as W. A. Tarr (1921, p. 374) was one of the first to note. He pointed to the distinctive physical relationships that exist between the shales and the concretions, evidence not available in the case of noncompactible sediments. He considered that it would be impossible after lith-

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Plate 3.—CARBONATE CONCRETIONS

Figure 1.—Embedded in dark-gray shale. Note the curvature of the shale bedding around the concretion. Courtesy Donald J. Podesta, International Petroleum (Colombia) Limited.

Figure 2.—In Cretaceous La Luna formation near La Vega, Cundinamarca. Note the apparent compaction or diminution of deposition below as well as above the concretion. Courtesy M. Dean Williams, International Petroleum (Colombia) Limited.

Plate 4.—CARBONATE CONCRETIONS

Figure 1.—In Cretaceous La Luna shales, Colombia. Note the apparent compaction of the shales below as well as above the concretion. Courtesy William D. Meyers, International Petroleum (Colombia) Limited.

Figure 2.—Post-Pleistocene *Mallotus* fish-bearing concretion. Photographed from specimen supplied by courtesy of the American Museum of Natural History.

ification of the shales to aggregate the calcareous material into a concretion.

Fish-bearing concretions are being found in Recent or sub-Recent marine clays along the coasts of Greenland and northern Canada (Pl. 4, fig. 2). These clays apparently were raised above sea level by isostatic rebound following melting of the Pleistocene ice cap. It is significant that concretions have already been developed in these very young clays. Probably concretion development is just one of many diagenetic processes that occur in sediments, organic as well as inorganic, during the period of physical, chemical, biochemical, and mineralogic instability upon deposition in radically new environments. The evolution of petroleum is thought to be one of these many diagenetic processes. As stability becomes established these changes cease.

That most of the fossil fish are full-bodied indicates that calcification and concretion growth were early enough to withstand compression from overburden, amounting in many places to thousands of pounds per square inch. For instance, the sediments containing the Colombia specimen were once buried beneath 10,000–15,000 feet of younger Cretaceous and Tertiary rocks, exerting pressures of at least 10,000 pounds per square inch. The manner in which the shale strata bend or "flow" around the concretions also indicates early concretion development, and subsequent compaction of the soft fluid-bearing muds around them.

These facts and the excellent state of preservation of the fossils, both plant and animal, all indicate a syngenetic origin. Lesquereaux (1870, p. 481–483) described concretions in Pennsylvanian shales in Grundy County, Illinois, where carbonate of iron concentrically developed around plants, bones of fishes, the remains of insects, and crustacea. The size and form of the concretions corresponded with those of the enclosed body. This is true of the Recent fish-bearing concretions of Canada and Greenland in which the physical structure of the fish is so beautifully preserved. Some of these specimens look as if the remains of the fish had been rolled in a batter, which had subsequently lithified and retained the form and dimensions of the fish.

Lesquereaux (1870, p. 482) emphasized the perfect preservation of the most delicate parts of the plant or animal, "just as if they were in life, or as if they had been transformed into stone while still living." He remarked on the absence of any flattening, attributing it to the early formation of the concretion. David White also supported the theory of syngenetic origin.

Origin of Carbonate Concretions

The manner of origin for all types of concretions is not easily explained, since carbonate concretions develop on a wide scale mainly in those facies in which limestones are not present or occur in very impure form.

The concretions under discussion are characteristic of a stagnant water environment in which oxygen is quickly used up, carbon dioxide accumulates, and lime is retained largely in solution as bicarbonate. Such environments today are usually of a low oxidation-reduction potential (E_h). They contrast noticeably with the carbon dioxide-free environments of shallow, aerated bottoms, such as those of the basin shelves and more limited bottom highs, where excess calcium carbonate is most readily removed from the waters and where the purest limestones usually accumulate. In fact, the preservation of organic matter indicates a reducing environment or its development soon after burial. The very presence of organic matter would tend to decrease the E_h.

There is no direct relationship between calcium carbonate deposition and the oxidation-reduction potential; it is primarily an environment association. The potential of a stagnant environment ordinarily ranges from low positive (only slightly oxidizing) to negative or reducing, but it is the stagnancy rather than the low O/R potential that holds lime in solution.[3]

[3] The oxidation-reduction may, however, have an important influence on the composition of some types of concretions and the surrounding relationships. Dr. Allen Tester (Personal communication) stated that "the development of iron oxide and iron carbonate concretions in Pennsylvanian and Cretaceous rocks is very common around plant and animal remains in sandy and silty beds with intimate preservation of the organic structures. In these cases the iron is characteristically reduced from limonite to goethite or hematite, and siderite is the typical form of the carbonate. The influence of the reducing action presumably caused by consumption of oxygen

The deposition of calcium carbonate is controlled mainly by the hydrogen-ion concentration (pH). Any condition, such as increase of temperature, agitation and aeration, or decrease of CO_2 pressure in the atmosphere, that causes removal of carbon dioxide increases the pH and so favors deposition of lime. Conversely, any decrease in alkalinity tends to hold the lime in solution. Sea waters are normally alkaline, with a pH ranging from about 7.5 to 8.5. A neutral solution has a pH of 7.0 and decreasingly lower values represent increasing acidity. Normally, calcium carbonate deposition requires a pH of at least 7.8.[4] On stagnant bottoms the pH would normally be well below 7.8 or even in the acid range. Why is it, then, that in the same environment, generally unfavorable for pure limestone deposition, such as that of the Cretaceous shales of Colombia, calcium carbonate concretions can develop locally?

There are ways in which the pH of an unfavorable environment may be locally increased to the point of carbonate deposition. Decomposition, if only anaerobic, goes on in most stagnant bottom environments. With anaerobic decomposition, ammonia or amines is locally concentrated. Bacteriologists use this test to determine if the bacterial process is progressing. Release of ammonia would increase the pH sufficiently to permit the precipitation of the bicarbonate as carbonate.

Once the initial concentration and precipitation of calcite crystals (seeding) start, they continue probably for some time. Seeding might be effected by some form of carbonate, such as a fragment of aragonite from a shell. Obviously the tissues of the fish must have been rapidly calcified. At the same time lime was thrown out of solution as the carbonate peripheral to the fish as long as any source of alkalinity persisted.

Bearing on the problem are some experimental observations by Franz Hecht (1933, p. 165–249) on what happens to the carcasses of animal organisms on the sea bottom or after shallow burial in the sediments.

Hecht studied the body of a shark found in 1924 near the outer edge of the largest tidal flat at Jade Bay near Wilhelmshaven, north Germany (Hecht, 1933, p. 197). The carcass was buried beneath a meter of sand at that locality, and left for two years. When it was uncovered in 1926 it had not changed noticeably in form. The specimen was again examined in 1931. Parallel studies of dead animal organisms were made in aquaria in the laboratory to check those on the near-shore sea bottom.

It was found that the albuminous or proteinaceous (nitrogen-bearing) substances decomposed much faster than the fatty constituents. The nitrogen is released in the form of ammonia or amines, making possible a degree of local alkalinity favorable to the deposition of calcium carbonate. Nitrogen escapes much more rapidly as ammonia than does carbon as CO_2, which accounts for the greater ratio of carbon to nitrogen in sedimentary organic matter than in living bodies (Hecht, 1933, p. 165–249; Trask, 1937, p. 150–154).

Hecht also reported that the end product of retarded decomposition is a homogeneous solid substance, perhaps adipocere which is a waxy brownish substance generated in dead bodies that have been long buried or exposed to moisture, especially sea water (Hecht, 1933, p. 201–205). Hecht observed that the more resistant fatty substances of the animal body may penetrate the enveloping sediment and

in organic decomposition can be seen to grade outward from the concretion. Thus a concretion of iron carbonate in a sand (or hematite) grades outward into limonite.' 'Dr. Paul Kerr of Columbia University relates (Personal communication) that the results of some recent mineralogic studies he has made of shales and other sediments indicate to him that siderite is a much more common constituent of sediments and also of nodules and other features of sediments than generally believed.

[4] The removal of silica from the lands and the marine deposition of chert from silica-bearing waters depends largely on the pH of the environment. Sea waters contain very slight amounts of silica, since the large quantities carried to the seas are quickly precipitated. Adjacent vulcanism, both submarine and subaerial, has been an important source of silica at times over the world. Also, in periods of extensive peneplanation dissolved silica comprises much of the diminished quantities of transported materials, and silica precipitation as well as lime and mud deposition is concentrated. This fact and the physical chemistry involved partially explain the common association of chert with the lime and mud facies which occurs in characteristic positions with respect to the basin-bottom form and environment that control the deposition.

cement it, concretionlike, as happened in the case of the shark.

Hecht observed that the carbon dioxide evolved during organic decomposition had a marked solvent effect on shells and bones (Hecht, 1933, p. 242). It probably explains why in certain facies of sediments only the impressions or molds of the shells are left. Upon decomposition, limestone-building organisms revert to their original stable products, carbon dioxide and water. This carbon dioxide, which is released to the solutions permeating newly deposited carbonate sediments, causes many of the distinctive features, such as vug to pinpoint porosity, great permeability, solution and recrystallization, and early forms of calcite veining characteristic of carbonate rocks formed in certain environments of basins. They are indicative of a well-defined environmental range of pH and Eh, degree of circulation or stagnancy, origin and rate of deposition (Weeks, 1952, p. 2122–2124).

Origin of Lime Impregnation of Muds

The Cretaceous basins of the Andean geosyncline are typical of many in which limestones and shales predominate. Environment determines the basin facies, and shales are characteristic of the deeper stagnant environment. Depending on the concentration balance of the respective ions, these shales may be slightly to very limy. Organic matter, which is completely decomposed in an aerobic environment, normally is not oxidized or may be only partially decomposed in the quiet stagnant environs of a basin. The resulting sediments usually are, or once were, bituminous. The extent to which they remain bituminous may depend upon the degree to which the clay or lime muds have given up their bitumen as hydrocarbons, or upon how much these fluid hydrocarbons have shifted to more porous facies. Widespread study of oil occurrence demonstrates that there is no relationship between the oil productivity of a series or basin and their content of highly bituminous beds. Some of the most prolific basins may carry little or none of these beds; conversely, basins or series rich in such beds may carry little or no commercial oil.

If concretion formation in unfavorable environments can be explained by local alkalinity around a piece of organic matter, perhaps lime impregnation of shales deposited in stagnant lime-bearing waters may be similarly explained. The limited decomposition, mainly anaerobic, of the organic matter may create a pH sufficiently high to permit release of calcium carbonate from the lime-bearing solutions permeating the muds. The resultant lime content would probably depend on the ion concentration and balance and on the supply of noncarbonate mud.

Though this paper may not have solved the problems posed, perhaps it may have aided in their solution.

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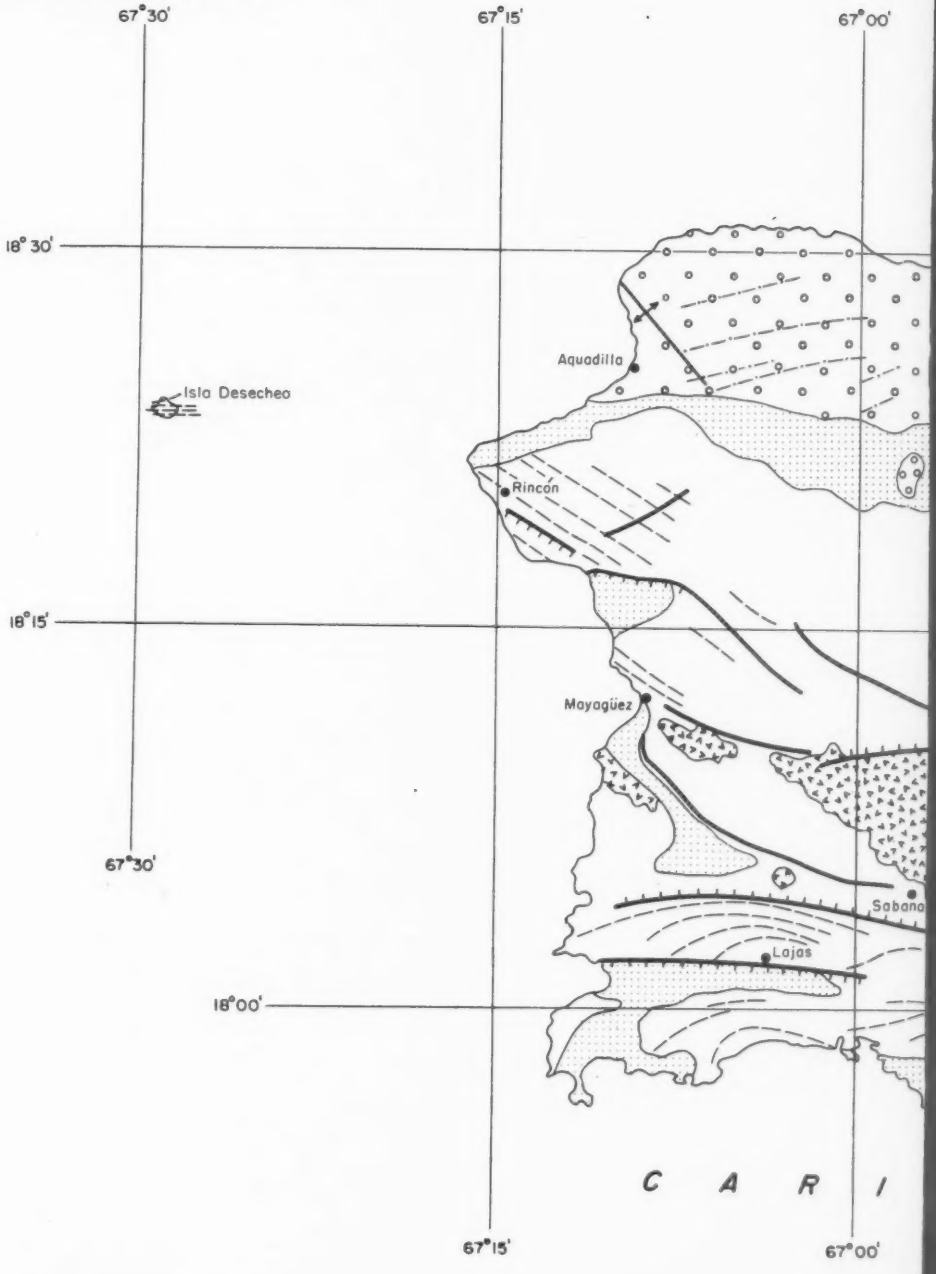
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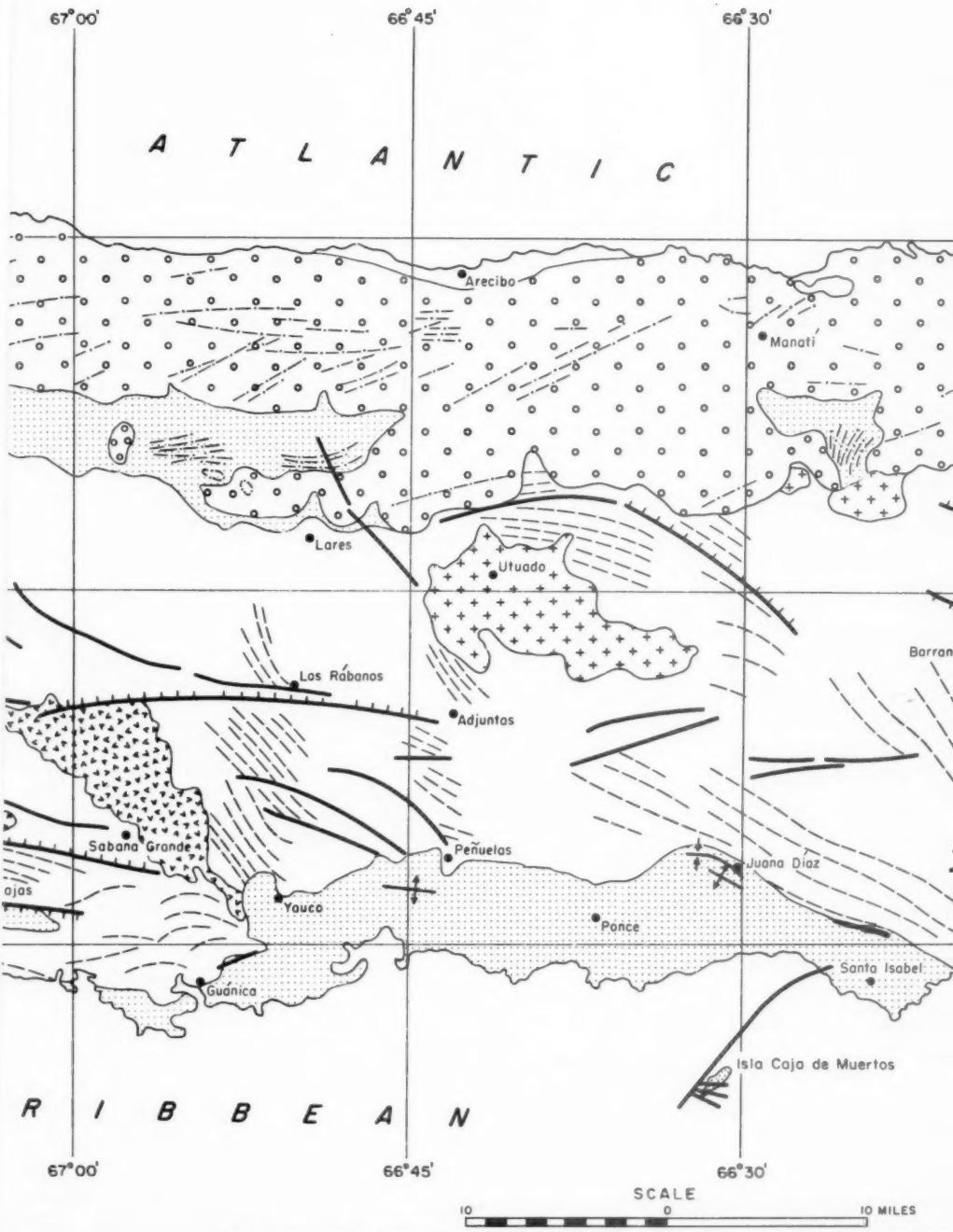

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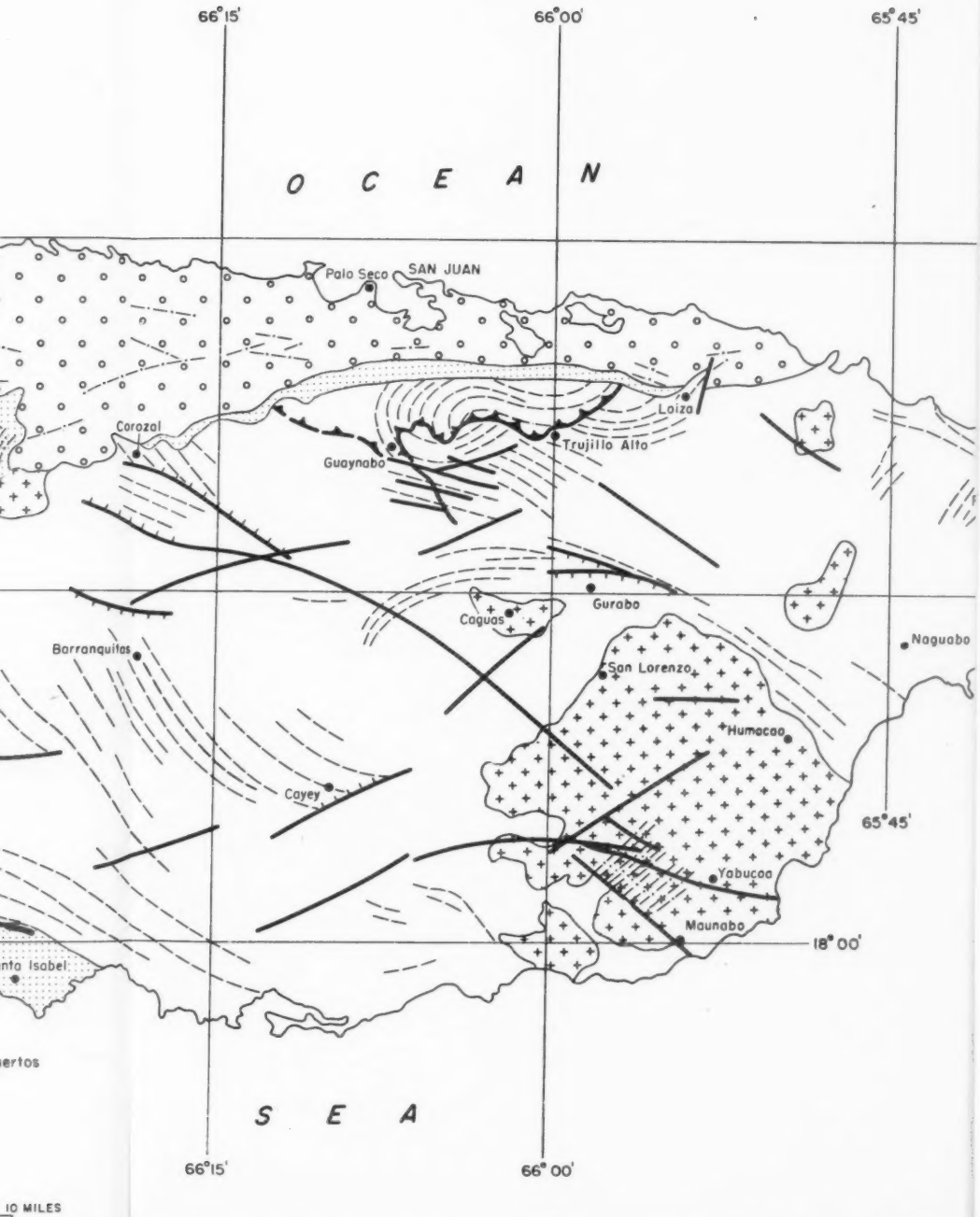

INFERRED STRUCTURAL FEATURES OF PUERTO RICO

EXPLANATION

Late Oligocene to late Miocene

Middle Tertiary rocks; dotted where predominantly marly limestone, or shale, sandstone and conglomerate; circled where predominantly dense limestone ("reef" facies) giving rise to prominent karst topography. (Source: Kaye, field and photo interpretation; Zapp et al., 1948, McGuiness, 1947.)

TERTIARY

Late Cretaceous to early Eocene(?)

Major older complex plutons
(Source: Meyerhoff: 1933; Roberts et al., 1942; Kaye.)

Older complex, serpentine
(Source: Roberts et al., 1942.)

Older complex, undifferentiated

CRETACEOUS and TERTIARY

Generalized strike trends of older complex
(Source: Kaye, field and photo interpretation.)

Thrust fault
(Source: Kaye, field.)
Dashed where inferred

Large faults and crushed zones, possessing for the most part topographic expression.
(Source: Kaye, field and photo interpretation.)

Prominent scarps that are probably also high-angle faults.
(Source: Kaye, field and photo interpretation.)

Jointing having prominent topographic expression

Anticline in middle Tertiary rocks

Syncline in middle Tertiary rocks

RICO

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 103-118, 2 FIGS., 5 PLS. JANUARY 1957

NOTES ON THE STRUCTURAL GEOLOGY OF PUERTO RICO

BY CLIFFORD A. KAYE

ABSTRACT

Two major structural and stratigraphic rock units occur in Puerto Rico: the older complex, ranging in known age from Late Cretaceous to late Paleocene or early Eocene and the middle Tertiary sequence, ranging from late Oligocene possibly to late Miocene. The former rocks are eugeosynclinal in character and are very badly faulted but for the most part apparently only moderately folded. With the exception of a large, partly low-angle thrust, the writer has not recognized evidence of strong tangential stresses. Intraformational folding in the older complex is interpreted as caused by submarine sliding (slump structure). Except in the vicinity of the larger plutonic intrusions and in the northeastern and southwestern corners of the island, the main strike alignment of the older complex is northwestward. The plutons are roughly concordant with the structure of the country rock and show varying degrees of differentiation. Their average composition seems to be more acidic than that of the volcanic and volcanogenetic rocks into which they are intruded.

The middle Tertiary sequence is nonvolcanic, made up dominantly of calcareous marine sediments. It crops out on the north and south sides of the island and in structural troughs on the west coast. On the north coast the beds dip gently to the north, and, except for slight terracings and a flexure at the northwestern corner of the island, are not folded. The middle Tertiary sequence on the south side of the island is somewhat folded. Seismic-reflection studies of the north coast indicate, however, a pronounced northward thickening, possibly some folding, and unconformities at depth. Unconformities which may be local have also been noted at several places on the surface. Several large faults in the middle Tertiary sequence have been recognized in both the north- and south-coast belts. The pattern of master joints that is inferred in the north-coast middle Tertiary belt from topographic alignments seems to indicate (1) control by facies contacts, (2) possible downwarping associated with the sinking of an arm of the Puerto Rican Trench, and (3) tension during upwarping along the island axis.

The dates of the major diastrophic events that are decipherable from the Puerto Rican data are: (1) early Tertiary (possibly late Paleocene, but more probably Eocene), when the older complex was deformed; (2) Miocene, when the middle Tertiary sequence was deformed; (3) late Pliocene and possibly early Pleistocene, when block faulting on a large scale produced the present topographic relief. Remnants of the fault scarps resulting from the latter deformation occur widely. Quaternary marine deposits and marine terraces suggest that Puerto Rico has been relatively unaffected by crustal movement at least since the late Pleistocene.

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Introduction

The following observations on some Puerto Rican structural features were collected during the writer's residence on the island from 1949 to 1951. As geologic consultant on engineering projects sponsored by several agencies of the Insular Government, the writer examined a variety of construction sites. In addition to these engineering geological studies, detailed geologic mapping of an area centered about San Juan yielded structural information of some interest. Particularly valuable as a source of structural data were the tunnels and dams of the Puerto Rico Water Resources Authority, especially the excavations for the Caonillas Extension Project, near Utuado, and the Southwestern Puerto Rico Project, centered about Yauco (Pl. 1). These projects provided a rare opportunity to study long sections of well-exposed rock in a region otherwise characterized by poor outcrops and helped clarify the nature of a few of the essential geologic structures that previously could only be inferred from scattered surface data.

Photogeology has also been used, particularly in preparing Plate 1.[1] Several of the larger structural and lithologic features that are difficult to see on the ground are readily perceived from the "air." These include striking topographic alignments, variously interpreted as marking large faults, crushed zones, and master joints, as well as the outlines of two main lithologic facies of the north-coast middle Tertiary belt: the denser limestones with their pronounced karst topography, and the chalky and marly beds with a more normal subdued topography.

No attempt is made to extend the Puerto Rican data to a regional synthesis.

[1] Plate 1 was prepared in 1951 and field-checked in the summer of 1952. A somewhat similar map by R. C. Mitchell (1954, Fig. 9) first came to the writer's attention early in 1956, 4 months after this paper was submitted for publication.

Acknowledgments

Four agencies of the Puerto Rican government co-operated financially with the U. S. Geological Survey: the Puerto Rico Water Resources Authority (abbreviated P.R.W.R.A. here), the Puerto Rico Economic Development Administration, the Puerto Rico Aqueduct and Sewer Authority, and the Insular Department of the Interior.

Special acknowledgment is made to the Puerto Rico Water Resources Authority for permitting the publication of tunnel data and to the Puerto Rico Economic Development Administration for permission to publish results of the geophysical survey of the middle Tertiary belts. The Guaynabo Magnetic Observatory, U. S. Coast and Geodetic Survey, and particularly its chief, Paul Ledig, are to be thanked for making their seismic records available.

Acknowledgment is also made of the work of John C. Dunlap, the writer's predecessor in Puerto Rico, who mapped two of the five tunnels in the Utuado area; his notes have contributed to the over-all structural picture.

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The Older Complex

The rocks of Puerto Rico are divisible on structural, lithologic, and stratigraphic grounds into two units: the middle Tertiary sequence

and the older complex. The characteristics of both units have been described by Berkey (1915), Fettke (1924), Hodge (1920), Hubbard (1923), Meyerhoff (1933), Meyerhoff and Smith (1931), Mitchell (1922), Semmes (1919), and Zapp, *et al.* (1948). The older complex has a known age range of Late Cretaceous to late Paleocene and possibly early Eocene (Kaye, 1956); it may include rocks both older and younger than this. In general the Puerto Rican older complex is similar to rocks of the same age elsewhere in the Antilles.

The older complex crops out as a broad strip that extends east and west across Puerto Rico and includes most of the hilly and mountainous terrain (Pl. 1). Approximately 15,000 feet of sedimentary and volcanic rocks belonging to the older complex have been mapped by the writer in a small area south of San Juan. The maximum thickness is not known. The rocks are lithologically varied, although almost everywhere marked by volcanically derived components. Most of the volcanic rocks are andesites and basalts, now variously altered. A few areas of banded rhyolite occur, but acidic volcanic rocks are the exception. Although volcanic flows and rocks easily recognizable as of pyroclastic origin constitute less than half of the outcrop area of the older complex, many of the finer clastic rocks are essentially either crystal tuffs and fine ash or sedimentary rocks derived from them by reworking. Most of the arenaceous rocks are graywackes, showing poor sorting and a mineralogically varied composition (some detrital quartz is present almost everywhere). The rocks fit into the eugeosynclinal category, as defined by Stille (1941) and Kay (1951).

Pure organic limestones—*i.e.*, those rich in skeletal matter—occur in the older complex at many places throughout the island. These are lenticular, and some locally attain considerable thicknesses (such as an estimated 900 feet of Trujillo Alto limestone that crops out about a mile north of the town of the same name). Tuffaceous limestones and calcareous tuffs are widespread, although their surface expression is deceptively small owing to the tendency for the outcrop to be leached. Macrofossils are rare in the older complex, although microfossils, most of which are Foraminifera, are abundant in the calcareous rocks.

The whole sequence is characterized by a general lensing of lithologic units, and local unconformities and cut and fill structures are common (Pl. 2, fig. 1).

The depositional environment of the Puerto Rican older complex probably included insular as well as marine facies. The thick accumulations of massive flow rock, conglomerate, and tuffaceous sandstone with current bedding and abundant carbonized plant tissue (wood?) in the eastern part of the island may represent alluvial deposits on the flank or flanks of a former insular cone or cones. Most of the older complex, however, appears to have been deposited in a sea, studded with large and small volcanic vents, which was undergoing more or less continuous subsidence. The detrital quartz found in many graywackes may have been derived from rhyolitic volcanic rocks, or from older quartz-bearing rocks (a "basement" complex dating from the Early Cretaceous or older?) which do not crop out, as far as the writer knows, on the island today.

The older complex is intruded by rocks ranging in composition from peridotite (serpentine) to granite. The intrusive origin of the serpentine in the southwestern part of the island is suggested by the contact metamorphic effects on the adjacent rocks (Kaye, 1953).[2]

The larger plutonic bodies, such as those in the southeastern corner and central part of Puerto Rico (Pl. 1), are composite complexes containing rocks that range in composition from gabbro to granite. Fettke's (1924) map of the pluton in the southeastern corner of the island indicates about 85 per cent quartz diorite; the rest is granite and quartz monzonite with minor gabbro. The smaller pluton at Utuado is chiefly granodiorite and quartz monzonite, with less quartz diorite and gabbro. The relative importance of the main rock types in this pluton is indicated by the tabulation of footages (Table 1) penetrated by the tunnels of the Caonillas and Caonillas Extension Projects (P.R.W.R.A.). This set of five tunnels, which

[2] Professor H. H. Hess (Personal communication), from his study of surface exposures, questions an intrusive origin for these rocks. The tunnel exposures studied by the writer showed, however, a very uneven blocky contact more suggestive of intrusion than of an unconformity or of faulting. This hypothesis is supported by the alteration aureole in the adjacent porphyry.

extend from Río Adjuntas on the southwest to Río Caonillas on the northeast, cuts across the pluton from wall to wall (the tunnels do not, however, form a complete and continuous traverse).

The south edge of the Utuado pluton, where seen, is steeply dipping to vertical, but on the north side its contact dips gently northward beneath the steeply dipping, bedded marine tuff and graywackes. From cross-cutting relationships, the order of intrusion of the above rock types is (oldest to youngest): quartz monzonite-granodiorite, quartz diorite, and gabbro. This relationship can be seen in the bed of Río Caonillas about a mile below the dam, and in the cuts along the new road leading from the dam to the Power House.

TABLE 1.—ROCK TYPES IN THE TUNNELS OF THE CAONILLAS AND CAONILLAS EXTENSION PROJECTS

Tunnel	Granodiorite and quartz monzonite (feet)	Quartz diorite, quartz diorite porphyry, quartz porphyry (feet)	Gabbro (feet)
Caonillas Power	2300	700	2000
Jordan-Caonillas	3000	—	—
Viví-Power	4400	—	—
Pellejas-Viví	3300	1650	—
Adjuntas-Pellejas		100	—
Total footage	13,000	2450	2000
Percentage	75	14	11

The older complex is very much faulted and is variably, though rarely intensely, folded, with moderate dips (<45°) more common than steep dips.

THE MIDDLE TERTIARY SEQUENCE

Unconformably overlying the older complex are the relatively undisturbed strata of late Oligocene, following Woodring and Thompson's, (1949) twofold subdivision of the Oligocene, to early Miocene age (Zapp *et al.*, 1948). Rocks of Miocene age, described by McGuinness (1947) as the Guanajibo formation, which he tentatively considered to be upper Miocene, occur in the structural valleys of the west coast (Pl. 1). All these rocks can be termed the middle Tertiary sequence. The main mass of middle Tertiary rocks crops out in broad east-trending belts on the north and south sides of the island (Pl. 1). These have most recently been described by Zapp *et al.* (1948).

The middle Tertiary strata are predominantly limestones and marls and show no evidence of contemporaneous volcanic activity such as characterizes the older complex. Light-colored thick-bedded dense limestones[3] are particularly important in the middle Tertiary belt of the north coast (Pl. 1). They are characterized by a peculiar topography in which haystack hills are the dominant form (Meyerhoff, 1938), representing a mature stage in the karst erosion cycle Chalky to marly limestone and lesser thicknesses of clay, sand, and gravel make up the remainder of the middle Tertiary deposits of the north coastal belt, as well as the south- and west-coast deposits in their entirety. On the north coast the latter rocks interfinger in an interesting facies relationship with the dense limestones (Pl. 1). The west-coast middle Tertiary rocks are almost everywhere buried by valley alluvium and to the writer's knowledge crop out at only one place—on the north side of the Lajas Valley, a little over 3 miles west-southwest of Lajas. Well-cuttings and foundation borings supply almost all present knowledge of these rocks.

Compared to the older complex, the middle Tertiary beds are slightly deformed. On the north side of the island the beds strike generally

[3] These have been called reef limestones by most writers (*e.g.*, Hubbard, 1923; Zapp *et al.*, 1948). The term is possibly appropriate, although the limestones are well stratified and do not have the massive structure of typical modern reefs or Paleozoic bioherms. The massive appearance of many outcrops of these Puerto Rican dense limestones is due to a secondary "caliche" coating which obscures the stratification.

PLATE 2.—UNCONFORMITIES

FIGURE 1. Local small-scale unconformity in calcareous tuff, right-abutment excavation, Yauco dam

FIGURE 2. Unconformity in lower Miocene limestone, east side of Guánica Bay. Beds below are approximately horizontal; beds above dip south.

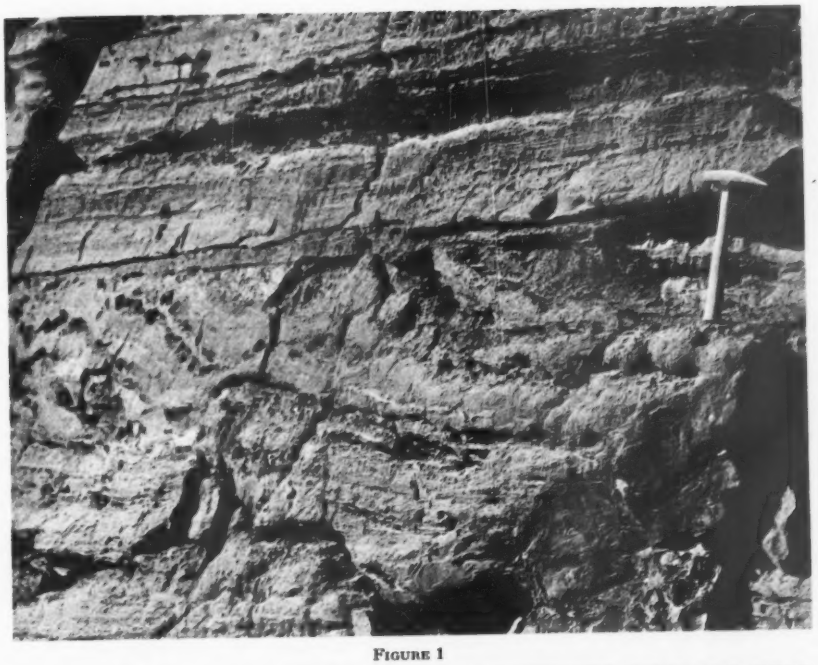


FIGURE 1

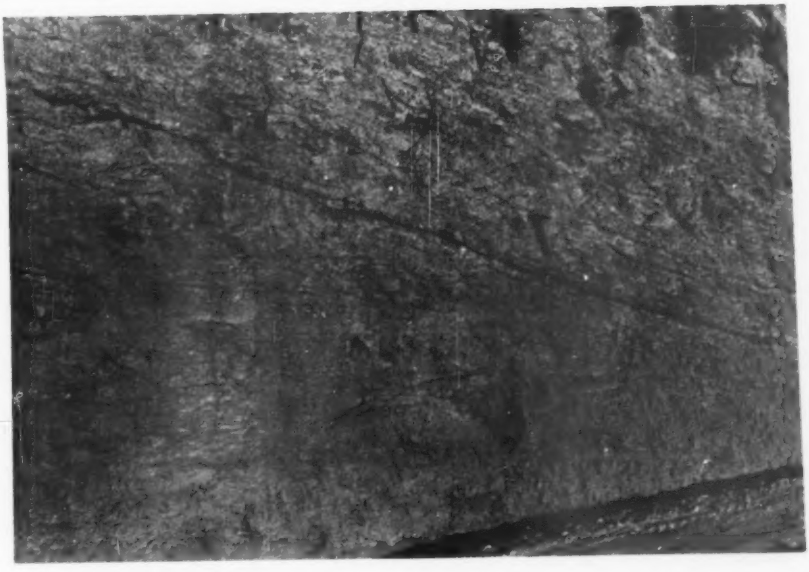


FIGURE 2

UNCONFORMITIES

FIGURE 1

FIGURE 2

STRUCTURES IN THE OLDER COMPLEX

east-west and dip gently northward. There is some terracing of this homoclinal dip but little surface evidence of folding. On the south coast the middle Tertiary sequence dips south, generally more streeply than on the north side, and is somewhat folded. It has been suggested (Lobeck, 1922; Hubbard, 1923), on the basis of faunal differences in the middle Tertiary rocks of the two coasts, that a thin ridge of land separated north-coast from south-coast deposition throughout late Oligocene and Miocene time when these sediments were laid down.

Outline of the Main Tectonic Events

It is clear that the major diastrophic event that affected Puerto Rico was that which terminated the deposition of the older complex in either late Paleocene or Eocene time. No evidence for a widespread unconformity within the sequence of older complex rocks was recognized by the writer (Kaye, 1956). Structural disturbance of the middle Tertiary rock is slight compared to the older complex. This diastrophism probably commenced in late Oligocene and culminated in the Miocene. Block faulting on a large scale apparently took place in the Pliocene and possibly early Pleistocene. It seems impossible to separate structures resulting entirely from block faulting from structures produced earlier.

Description of Structural Features

Intraformational Structures of the Older Complex

In two areas the writer saw tight folding that apparently resulted from flow of unconsolidated sediment. From their local development within rocks that otherwise show no evidence of tight folding, these folds seem clearly to be intraformational and nontectonic. That some, if not all, formed before consolidation was shown in the excavation for the right abutment of the Yauco dam, 3¾ miles north-northwest of Yauco. Here a thick blocky basaltic flow overlies thin-bedded tuffaceous limestones. The top 6 feet of the limestone is very contorted, doubtlessly because of the motion of the volcanic flow over its soft muddy bed.

Large-scale slump structures, consisting of grossly contorted shales and somewhat silicified ash, are well exposed in road cuts about 1½ miles southwest of Barranquitas (Pl. 3, fig. 1). Similar flow contortions were seen at several places in the well-bedded tuffaceous limestones and calcareous tuffs of the Prieto-Yauco tunnel. One of these, occurring about 8000 feet from the outlet, demonstrates how at least some of these structures originated by submarine sliding. Here a massive limestone breccia abuts against contorted, thin-bedded tuffaceous limestone (Fig. 1, top). The bedding of the limestone is wrapped around the end of the limestone breccia, and the breccia itself penetrates the limestone along bedding planes and fractures for several feet. The limestone breccia consists of fragments, generally less than 2 inches across, of a pure organic limestone embedded in a tuffaceous matrix. This organic limestone is unlike the adjacent contorted beds but has a bedded counterpart in several lenticular massive rudist-bearing limestones that crop out to the south. Limestone breccias like the one in the tunnel crop out in many places in the Yauco area where they constitute broad though thin lenticular masses. Kuenen and Migliorini (1950) describe similar breccias from the Apennines under the term "brecciolas" and attribute them to submarine sliding. The Puerto Rican limestone breccias are probably for the most part submarine slide debris, and the tunnel seems to expose a slide frozen while burying its nose (or its toe) in what must have been poorly consolidated limey muds. The displacement and rumpling of the muds by the slide and the incorporation of twisted slabs of partly consolidated mud in the edges of the slide debris are evident.

Plate 3.—STRUCTURES IN THE OLDER COMPLEX

Figure 1. Complex intraformational folding 1.6 miles southwest of Barranquitas. Stratification pointed up with black ink. A certain amount of displacement along horizontal shear planes is evident.

Figure 2. Fault-line scarp, south end of Isla Caja de Muertos. Thin-bedded, fine-grained rocks of older complex underlie grassy slope on right; middle Tertiary dense limestone underlies brush-covered slope on left.

A thin zone of contorted black shale (Fig. 1, bottom), resembling the contorted structures described by Shrock (1948, p. 156–161), Emery (1950), Natland and Kuenen (1951), Kuenen and Menard (1952), Kuenen and Carozzi (1953), and Kaye and Powers (1954) and sometimes called flow casts or load casts, was also noted in the Prieto-Yauco tunnel. Possibly these structures are caused by the foundering of sand into underlying soft muds, resulting in a squeezing up of the mud in complex intrusive forms along breaks in the foundering layer.

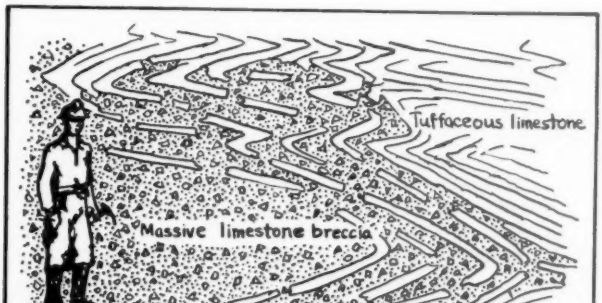


FIGURE 1. INTRAFORMATIONAL DISTORTION AT TWO PLACES IN PRIETO-YAUCO TUNNEL

Structures Resulting from the Early Tertiary Orogeny

Structural trends and folds.—The older complex is deformed into large folds whose orientations are indicated by the strike trends shown in Plate 1. These trends are based on the writer's field notes and do not cover the entire older complex outcrop. The strike trends shown are partly interpretative, for divergencies of strike that seem to be due to local folding and faulting have been eliminated. Strike trends rather than fold axes are shown, because without clear-cut stratigraphic markers and with much faulting the major folds of the older complex are difficult to delimit without detailed mapping. The importance of faults in obscuring folds in the Puerto Rican older complex cannot be overemphasized. Strongly divergent attitudes in adjacent outcrops cannot *a priori* be construed as an expression of folding. Rotation by displacement along curved fault surfaces, fault drag, and the juxtaposition of elements of different folds by faulting complicate the reconstruction of folded structures. Hence road-cut traverses do not suffice to outline folds.

Nevertheless, certain observations can be made about folding. The prevalence of northwesterly strikes is evident in Plate 1. Notable exceptions lie in the southwestern corner of the

island and near the large plutons at Utuado and in the southeastern corner of the island. The curvature of strikes about these large intrusive bodies expresses the essentially concordant nature of the plutonic injections (they generally are not concordant in detail). The festoon of curved strike trends in the southwest corner of the island suggests that a pluton may lie off the coast. A roughly comparable structural trend exists near San Juan, but close study has provided no evidence for an intrusive origin.

The general form of the major folds in the area lying immediately south of San Juan are outlined in Plate 1. This area was mapped in detail (Kaye, in preparation). A broad plunging anticline with an essentially east-trending axis and a fragmentary south limb is suggested here. It is not known whether this structure and the thrust that underlies it are genetically related. Dips generally do not exceed 45°, except where caused by fault drag.

The flanks of the larger folds are terraced and broken by smaller structures whose orientations commonly diverge from the axes of the major features. Several of the tunnels studied illustrate these secondary structures. Plate 4 illustrates two tunnels that showed the best-developed folds. The structures have been generalized to eliminate the effects of faulting. The folds are neither tight nor indicate a dominant tangential stress. Near its outlet portal the Adjuntas-Pellejas tunnel of the Caonillas Extension Project (P.R.W.A.) cuts the western edge of the Utuado pluton. The dip steepens at the intrusive contact. Several fold axes are suggested in this tunnel. The northwest-trending axis near the midpoint is probably a local flexure. A major synclinal axis is suggested about 2000 feet from the inlet portal, although the massive quartz andesite in this stretch of tunnel makes this interpretation difficult to verify. A large fault (judging from the amount of gouge) crosses the tunnel near here, and its effect on the local structure is unknown. Exposed along three stretches are tight drag folds in thin-bedded cherts that resulted in part from folding and possibly in part from faulting. Among these, the folds clearly caused by drag on the flanks of larger structures, have vertical to steeply dipping axial planes suggesting the essentially symmetrical form of the large folds. Small-scale current bedding and graded bedding exposed in the tunnel indicate that there was no overturning of beds, substantiating the interpretation of relatively open symmetrical folding.

Plate 4 also interprets the folds penetrated by the upper part of the Yauco-Loco tunnel of the Southwestern Puerto Rico Project (P.R.-W.R.A.). Several minor fold axes occur, although only one is closely parallel to the northwest regional strike. Most structures express poorly oriented warplike folds, typical of many smaller folds in the older complex.

Drag folds like those in the Adjuntas-Pellejas tunnel are common throughout the older complex in thin-bedded rocks and except for slump structures are almost the only clear-cut tight folds.

Faults.—Faults are abundant; from the area studied, the writer has the impression that the structural complexity of the older rocks is due more to faulting than to folding. Most sizable outcrops expose at least one fault, and the tunnels exposed many (Pl. 4). These range in dip from low angle to vertical. The displacement can rarely be assessed, as reliable stratigraphic markers are generally absent. Slickensides suggest highly varied movement along even closely spaced faults. Slickensided joint planes in the massive rocks, including the plutonic rocks, are common and demonstrate that differential movement has probably occurred along most fractures.

Detailed mapping by the writer of the older complex in the area around San Juan indicates the existence of a large thrust fault, trending essentially east-west and for part of its length inclined at a low angle. This fault was poorly exposed at two places; most of its trace was inferred from structure and stratigraphy. It apparently superimposes structures of different trends (Pl. 1). By rearranging the folds below and above the thrust into a continuous system, a displacement of at least 2 miles is indicated, with the upper plate moving south in relation to the lower. Most of the other faults shown in the San Juan area are steeply dipping. Many that trend northwest are marked by eroded scarps that may be remnants of true fault scarps resulting from crustal movements in the late Pliocene or early Pleistocene.

Shown on Plate 1 are certain lineations in the older complex that either have decided topographic expression or can be clearly read from aerial photographs. These lineations consist either of scarps or of rigidly aligned valley-and-col systems. They are thought to mark long high-angle faults or major shear zones. In several places north of Yauco this interpretation was checked in the field. One fault is exposed in the outlet portal of the Prieto-Yauco tunnel (P.R.W.R.A.) and can be traced on the surface in an arcuate course for several miles to the southeast. Apparently this fault has a large though undetermined stratigraphic displacement.

Occurring about 10 miles north of Yauco is a large fault that is topographically expressed for more than 18 miles. This fault is marked by a prominent north-facing scarp for part of its length and by valley-and-col alignment for the rest. The trace is almost a true circular arc with an 80-mile radius and a center of curvature due south of the midpoint of the curve.

Possibly some of the faults in the older complex date from the late Tertiary rather than the early Tertiary. No criteria are known for distinguishing the age of a fracture in the older complex or for recognizing whether a given displacement of apparent late Tertiary age may represent reactivation of an early Tertiary fault.

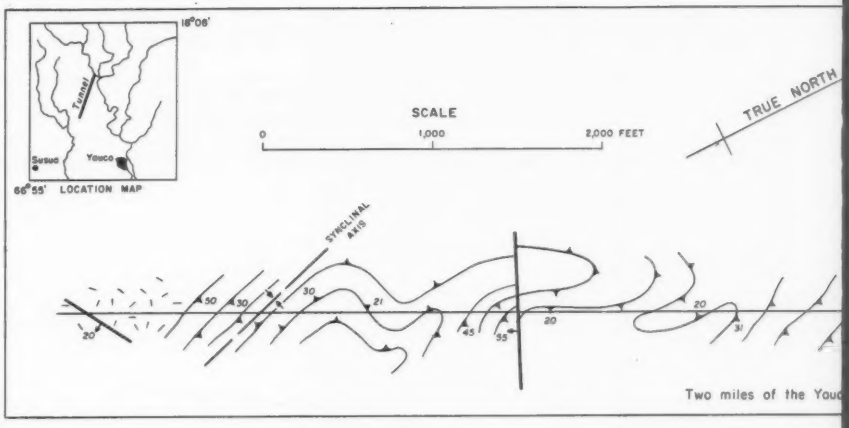
Middle Tertiary Structures

Unconformities.—Unconformities that are probably local have been seen at several places in the middle Tertiary sequence. The most obvious of these (Pl. 2, fig. 2) is exposed on the east side of Guánica Bay on the south coast in a road cut a few hundred yards south of the fertilizer plant. The stratigraphic significance of this unconformity is not known, and Zapp *et al.* (1948) do not mention it in their study of these rocks.

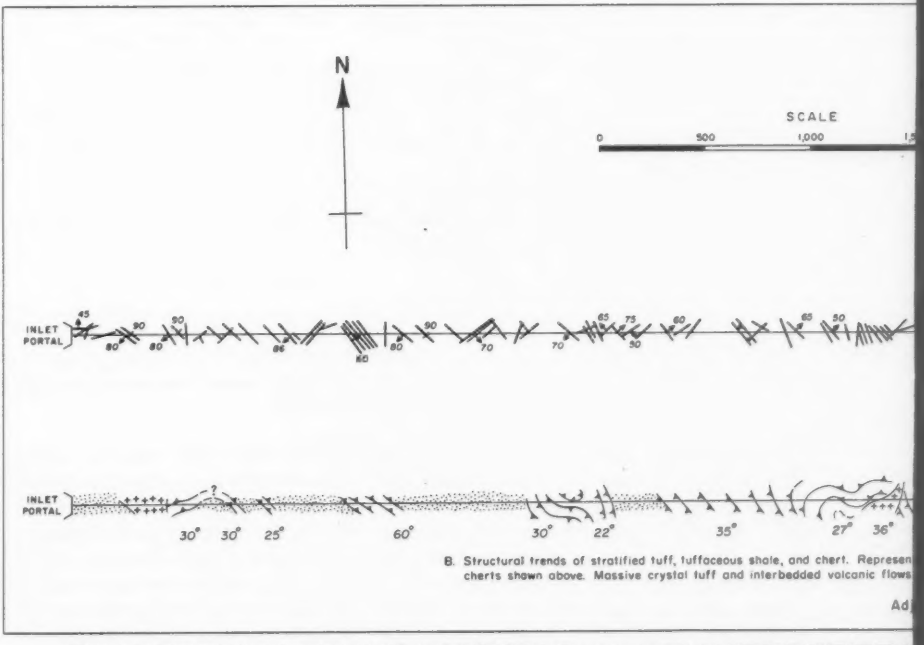
One and possibly more unconformities and a pronounced thickening of the middle Tertiary section are indicated by a seismic-reflection reconnaissance of the north-coast middle Tertiary belt made by the United Geophysical Company for the Puerto Rico Economic Development Administration (Myers, 1955). One of several cross sections made in this study is reproduced in Figure 2 and suggests a small unconformity at a depth of approximately 2900 feet. Presumably the reflecting horizons both below and above this break are middle Tertiary in age. The cross section is a mile long and is situated at Palo Seco, on the west side of San Juan Bay. Reflection data from stations between Arecibo and San Juan show that in this area the middle Tertiary section thickens markedly to the north. A maximum thickness of nearly 11,000 feet was recorded at a station about 8 miles west of Manatí and 2 miles south of the coast. This is nearly three times the thickness indicated by surface exposures at this meridian (Zapp *et al.*, 1948). These reflection data indicate that dips increase with depth and deviate markedly from the prevailing north regional dip of shallow horizons. These deeper rocks are probably separated from more shallow horizons by one or more unconformities.

Folding.—The map by Zapp *et al.* (1948) of the middle Tertiary belts shows how the rocks on the north side of the island dip homoclincally northward, whereas those on the south are downwarped more steeply to the south, defining a geanticlinal axis approximately coincident with the main east-trending topographic divide of the island. The gentle dips of the exposed north-coast beds rarely exceed 7°, whereas those of the south-coast belt commonly exceed 20°. Zapp *et al.* identified only one fold in the north-coast belt, a northwest-trending structure in the northwest corner of the island near Aguadilla (Pl. 1). The probable existence of buried folds in this belt, as shown by seismic-reflection data, has been mentioned. Surface studies by the writer indicated local terracing of the north-coast belt, although no south dips were found.

In contrast to the north-coast belt, folding on both a large and small scale occurs in the middle Tertiary rocks on the south side of the island. Zapp *et al.* (1948) recognized two major folds, one near Peñuelas and the other near Juana Díaz (Pl. 1). A detailed seismic-refraction study of the coastal strip east of Ponce (Denning, 1955) revealed general large-scale warping or folding (Pl. 5). Smaller structures can be seen in the cement quarry at Ponce where lower Miocene marls and limestones are deformed into a series of low undulations with amplitudes of 40 feet or more.



Two miles of the Yauc



B. Structural trends of stratified tuff, tuffaceous shale, and chert. Represen
cherts shown above. Massive crystal tuff and interbedded volcanic flows

Ad

TWO PUERTO RICAN TUNNELS, WITH INTERPRETA

EXPLANATION

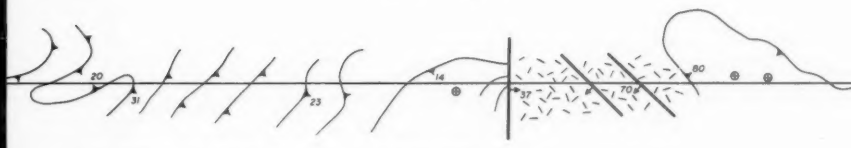
Upper Cretaceous–lower Eocene(?)

Rudistid limestone, limestone breccias and interbedded basaltic pyroclastics and flows. Fragments of limestone commonly included in volcanic rocks and fragments of volcanic rocks included in limestone.

Prevailingly massive basaltic flows and pyroclastics.

FEET

TRUE NORTH



Two miles of the Youco–Loco Tunnel. Only larger faults are shown.

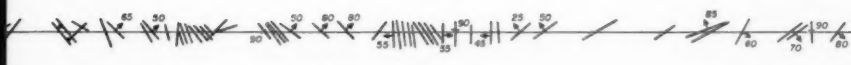
EXPLANATION

Strike and dip of faults

Strike trend and direction of dip of stratified rocks

SCALE

500 1,000 1,500 2,000 FEET

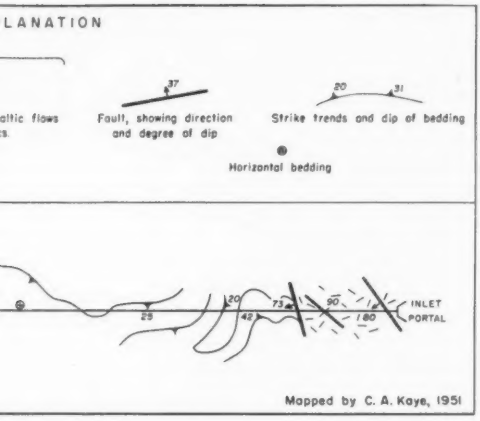


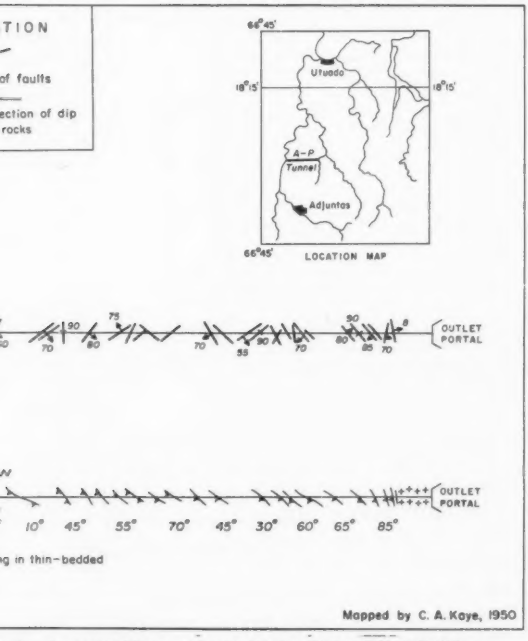
A. Faults

uffaceous shale, and chert. Representative dips shown below appropriate tunnel station, stretches of tight chevron folding in thin-bedded tuff and interbedded volcanic flows designated by dot pattern and intrusive rocks by cross-pattern.

Adjuntas–Pellejas Tunnel

INTERPRETATIONS OF STRUCTURAL TRENDS OF





OS OF THE OLDER COMPLEX

Faulting.—Although small normal faults are common in middle Tertiary rocks, large faults have been recognized or inferred in only a few places. Zapp *et al.* (1948) show only one fault, the floors of the several west-coast valleys and the older complex of the valley rims. These valleys probably have resulted from block faulting of either late Pliocene or Pleistocene

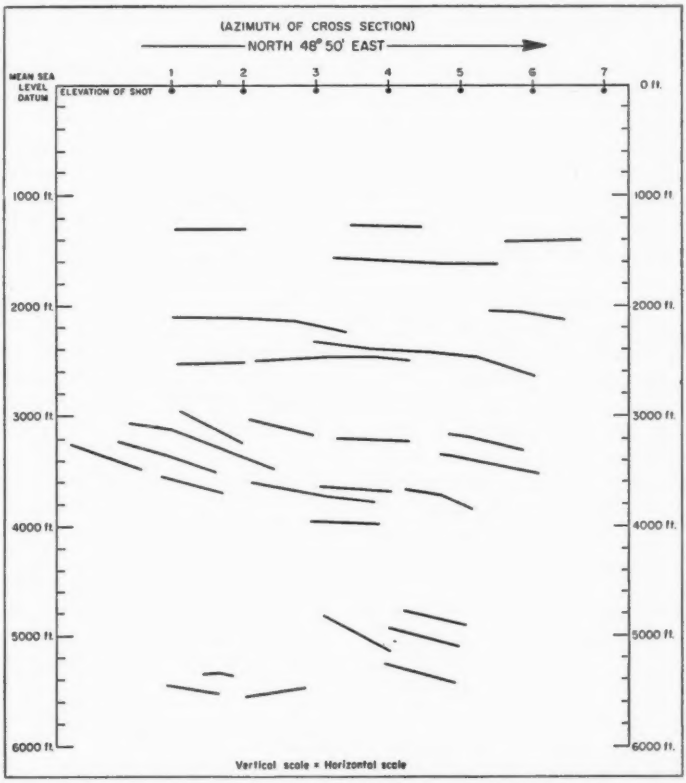


FIGURE 2. SEISMIC-REFLECTION SURVEY; VERTICAL CROSS SECTION AT PALO SECO, PUERTO RICO, SHOWING REFLECTING HORIZONS

east of Guánica, on their map. They inferred probable boundary faults separating the older complex from the middle Tertiary sequence at Juana Díaz and Peñuelas. Berkey (1915) and Mitchell (1922) believed that a south-coast boundary fault existed between the older complex outcrop and the middle Tertiary beds. As suggested by Lobeck (1922), however, the dissected nature of the unconformity between the older complex and the middle Tertiary of the south coasts makes the existence of this fault, or faults, difficult to determine. Nevertheless, there is good reason to postulate faulted contacts between the middle Tertiary deposits in date for it seems clear that they could not be topographic relics of Miocene depositional troughs, and no evidence exists in the surrounding rocks of the older complex that they are synclinical in origin.

The best evidence of a large fault in the middle Tertiary is provided by the seismic-refraction survey of the south-coast area east of Ponce (Denning, 1955) and by the rocks that crop out on the small Caja de Muertos island, a few miles off the coast (Pl. 5). A long, north-northwest-trending fault, which passed just west of the Isla Caja de Muertos, was indicated by the geophysical survey. This fault was

originally suggested as an explanation for the absence of energy return along a narrow zone in an area otherwise characterized by excellent refractive properties. The existence of this fault is given strong support by the three large faults that crop out on the Isla Caja de Muertos. This small island is made up of four separate fault blocks: two of clastic beds and limestone of Miocene age, alternating with two of bedded ashy shale and volcanic breccia of the older complex (Pl. 5, inset). Figure 2 of Plate 3 is a view to the east at the south end of the island and shows the southernmost of the three faults; it separates rocks of the older complex on the right from dense limestone of Miocene age on the left.

The probable stratigraphic displacement along each of the three Caja de Muertos faults is estimated to be at least 2000 feet. The basis for this figure is that older complex rocks are juxtaposed against rocks of Miocene age. Zapp *et al.* (1948) indicate a thickness of 2000 feet for the upper Oligocene cropping out on the mainland 10 miles due north of the island. If the same thickness of upper Oligocene underlies the lower Miocene beds cropping out on Isla Caja de Muertos, a minimum stratigraphic displacement of 2000 feet is indicated. Substantial support for this figure is given by the structural contours of the seismic-refraction map (Pl. 5).

The exposures on Isla Caja de Muertos, in conjunction with the seismic-refraction interpretation, convey a picture of a broad northeast-trending fault zone with at least several large auxiliary cross faults. Thus, the Isla Caja de Muertos is possibly a sliver—or the remnant of a sliver—of this fault zone that was squeezed or thrust up during late Tertiary or early Quaternary block faulting. The original fault dislocations, however, may date from the Miocene disturbances that ended the deposition of the middle Tertiary sequence.

At two places on the north side of Puerto Rico, faulting is inferred to involve both the older complex and the middle Tertiary. The easternmost dislocating is about a mile north of Loíza, where rocks of the older complex are exposed in a quarry in the western end of a typical haystack hill of middle Tertiary limestone. Although the contact of the older complex and the middle Tertiary rocks, which make up most of the hill, is not exposed, there seems little doubt that it is a fault; in Plate 1 this is shown as a continuation of a fault which the writer thinks passes up the valley of the Río Canóvanas to the south.

A north-northwest-trending topographic groove, apparent on aerial photographs, cuts both older complex and middle Tertiary rocks near Lares (Pl. 1). In its middle Tertiary sector the stratigraphic displacement is apparently small.

Jointing.—Striking topographic alignments, most of which are rectilinear and best seen on aerial photographs, occur widely in the north-coast middle Tertiary belt. In the marly and chalky facies these form furrows and grooves. In the dense limestones most are positive features consisting of alignments of the haystack hills. In the field the erosional features are seen to be enlarged joints and zones of close jointing. The positive topographic features in the dense limestone are also considered to be caused by joints.

Flint *et al.* (1953) referred to casehardening of friable limestone to explain the origin of certain narrow limestone ridges on Okinawa. Essentially, the casehardening process involves the deposition of secondary $CaCO_3$ by evaporation in the pore spaces of a relatively porous limestone. This takes place adjacent to surfaces exposed to alternate wetting and drying, particularly steeply inclined surfaces, and the effect is most pronounced in canyon walls and the sides of enlarged joints. This deposit of interstitial secondary lime forms a superficial crust, or armor plate, of dense limestone. Where residual soils can accumulate, as on the surface of a limestone plateau, evaporation is inhibited by the soil blanket, and casehardening does not occur. Here, the rate of topographic reduction brought about by limestone solution exceeds that on the narrow casehardened zones which characterize the steeply dipping surfaces. Narrow ridges or walls of dense casehardened limestone, such as those Flint *et al.* (1953) describe from the Pacific, are progressively raised in relief as the upland surface is reduced. On Puerto Rico, a convincing example of this geomorphic form occurs along the lower reaches of the canyon of the Río Guajataca, on the Puerto Rican north coast, where a narrow rampart or wall, 100 feet high, rises above the upland from the narrow canyon rim.

By analogy with these features, the ridges and aligned knobs in the denser limestone facies of the north-coast middle Tertiary belt are thought to be the densified walls of enlarged joints raised in relief by the reduction of the surrounding more porous limestone. This hypothesis cannot be proved in the field except by extensive drilling, but it is tenable and is the only hypothesis known to the writer that explains the curious alignments of knobs and ridges.

Joints in the marly and chalky limestones tend to conform to the contact between the dense and chalky limestone facies (Pl. 1). The spacial distribution of these facies seems relatively complex, particularly in three dimensions. A radial contraction of the marly limestone, away from its contact with the dense limestone, might produce a joint system similar to that seen on the aerial photographs and indicated in Plate 1. The interrelationship between joint pattern and facies outline may, however, involve factors other than simple contraction.

Several large annular depressions, probably also a reflection of jointing, occur from 4 to 6 miles northwest of Lares, near the margin of the dense limestone facies (Pl. 1). These structures, too, are possibly explained by contraction away from the highly uneven surface of the facies contact. A more plausible cause for these annular fractures, however, is differential compaction of the middle Tertiary sediments over buried hills. One such hill is exposed at the margin of the middle Tertiary belt, 3 miles northwest of Utuado and about 10 miles east-southeast of the annular depressions. This hill of older complex rock projects at least 300 feet above the general level of the older complex-middle Tertiary unconformity.

In the dense limestone facies, the trends of the inferred joints are less varied than in the chalky rocks (Pl. 1). Most of them are rectilinear and are oriented both northeast and east, rarely northwest. Except for the northeast-trending joints near Utuado, the joints tend to conform to the shore line. Possibly these are tension fractures formed during the subsidence of the Aguadilla re-entrant of the Puerto Rican Trench (*see* H. O. Chart 5487) or during the uparching of Puerto Rico in the Miocene.

The northeast-trending joints in the northwest corner of Puerto Rico approximately parallel the bathymetric contours that outline the southeast side of the Aguadilla re-entrant, an arm of the Puerto Rican Trench that heads southward into the Mona Passage. The foundering of this oceanic deep may have dragged down a portion of the crust lying to the south and east, resulting in northeast-trending tensional fractures.

The prevailingly east-west joints that tend to parallel the coast may have formed later as a result of arching in the later Tertiary and early Quaternary. If this is true, they demonstrate the close connection between the general outline of the island's north coast and crustal arching and emphasize the essentially tectonic origin of the island's present shape.

Late Tertiary Structures

The present geomorphic cycle of Puerto Rico probably dates from late Pliocene or early Pleistocene time. As noted by many writers on Puerto Rican geology (*e.g.*, Lobeck, 1922; Meyerhoff, 1933), the accordant summits of the island's high interior strongly suggest an uplifted peneplain or old-age surface. This surface, tracts of which are apparently still extant (for example, the low rolling uplands around Aibonito), must have been deformed in very late Pliocene or early Pleistocene. The uplift took the form of arching with attendant block faulting about an axis conforming approximately to today's east-trending insular divide. Important in this uplift were fault displacements, most of which were probably along breaks dating from the early Tertiary orogeny. Traces of this faulting can still be found in the present topography where the many abrupt changes in general upland levels, common north of the insular divide, probably represent remnants of late Tertiary or early Pleistocene fault scarps. Some of the more striking of these eroded fault scarps are indicated in Plate 1. Not all of them north of the island divide face north, however, and evidently the pattern of block faulting did not adhere to a simple uparching of the geanticlinal island. Some fault blocks evidently lagged behind the general arching and resulted in graben valleys and south-facing escarpments, such as that about 8 miles south of Corozal. Most of the grabens are at the eastern and western ends of the island

and comprise low-lying alluvial-filled troughs and embayments. On the west coast these valleys include, from north to south, the lower Añasco, the Guanajibo, and the Lajas. On the east coast the broad ramifying Naguabo-Humacao-Caguas lowland is probably a complex of blocks relatively unaffected by the late Tertiary elevation. The two triangular coastal plains on which are situated the towns of Yabucoa and Maunabo, in the southeast corner of the island, preserve most clearly signs of their faulted origin. The high dioritic ridge, Cuchilla de Panduras, which separates the two basins, is bounded by high-angle, nearly rectilinear faults that trend northwestward. The deep submarine embayment close off shore at this corner of the island (U.S.C.G.S. Chart 920) may be due to movement along these same fault blocks. The alluviated basin in which the city of Cayey is situated is clearly bounded on the south by a straight steep fault scarp and provides possibly the only sizable late Tertiary graben in the interior of the island.

All the scarps shown in Plate 1 probably date from late Pliocene or early Pleistocene, although as already suggested the original fractures may be much older. Because the large-scale faulting in the late Tertiary makes it difficult to assign a specific age to any fault on the island, the description of faults under the categories of early and middle Tertiary origin must be considered tentative; many of them may date from the late Tertiary.

A study of shore-line features of Puerto Rico (Kaye, in preparation) indicates that the coastal areas have been stable, in so far as vertical crustal movement is concerned, since the late Pleistocene. This conclusion is based on varied evidence, including: (1) the occurrence of well-developed sea-level-solution phenomena in carbonate rocks, which can be tied with some precision to modern sea level; (2) for some stretches of coast, clear-cut indications of a long history of shore progradation whose relationship to sea level can be determined; (3) in many places the occurrence of traces of several sea-level stands (presumably eustatic) which correlate with those reported from many parts of the world.

Seismicity

Puerto Rico is an area of moderate seismicity. The severest earthquake in modern times was on November 11, 1918, and apparently centered in the northern part of the Mona Passage (the Aguadilla re-entrant). This earthquake has been well documented by Reid and Taber (1919), who also list historic records of earlier Puerto Rican shocks. Rutten and Van Raadshooven (1940) tubulated earthquake epicenters in the Caribbean region from 1913 to 1939. They show only 2 epicenters on, or very close to, Puerto Rico, 10 epicenters in the Mona Passage and the Aguadilla re-entrant, and 4 epicenters close to the axis of the Puerto Rican Trench. During this interval greater seismic activity centered on the Dominican Republic and Haiti than on Puerto Rico.

The writer plotted the epicenters of earthquakes from 1941 to 1953, using as a source "Preliminary determinations of epicenters" (U. S. Coast Geod. Survey) and records in the U. S. Coast Geod. Survey Magnetic Observatory at Guaynabo. During this period 3 epicenters probably marked Puerto Rico proper, 2 of these approximately on the north coast, and 1 somewhere in the interior of the island. In contrast to the 1913–1939 period, no epicenters were reported from the Mona Passage. Four epicenters were located 30–50 miles north of the island, and 4 were close to the axis of the Puerto Rican Trench. Earthquakes centering about northern Dominican Republic were more frequent and, in general, more severe, as they were in the period 1913–1939.

Instrumentally determined epicenters do not give the entire picture of seismic activity of Puerto Rico. Reports of tremors felt on Puerto Rico have been tabulated from published reports (*Monthly Weather Review;* U. S. Coast Geod. Survey seismic publications) for the period 1919–1939. These include many feeble tremors of apparently local origin whose epicenters have not been determined because there is no instrumental data. They show an average of 3.4 sensible shocks a year, varying in frequency from 1 to 10 shocks per year. Recently installed seismographs at the Guaynabo observatory have registered many feeble shocks of probable local origin. From February,

1949, to April, 1951, 97 tremors of this type were recorded.

While these data show that the crust beneath Puerto Rico is still undergoing tectonic adjustment, most present major-crustal rupturing seems to be centered under the Puerto Rican Trench and its westward structural prolongation in the north-coast region of the Dominican Republic.

Summary and Conclusions

The scattered data now available do not reveal an unequivocal outline of Puerto Rico's geologic past. There is still too little systematic ground study to provide a sound basis for synthesis. The following summarizes the observations and conclusions at present justifiable.

(1) The known age range of the Puerto Rican older complex is Late Cretaceous to late Paleocene or possibly early Eocene. Older and younger rocks may also be present.

(2) No island-wide unconformity within the older complex is known to the writer. At Loíza, the one locality where undoubted upper Paleocene or lower Eocene rocks can be seen in contact with Upper Cretaceous rocks, the relationship seems conformable (Kaye, 1956).

(3) The older complex, though principally marine, probably included some subaerial deposition and was laid down in a region of intense volcanic activity and crustal downwarping. Apparently little if any difference marked the depositional environment or source material in early Tertiary and Late Cretaceous time, for the rocks of both ages are essentially the same.

(4) The presence of detrital quartz in the older complex shows that sialic crustal material was available. The quartz either came from quartz-bearing volcanic rocks or older quartzose rocks (intrusives ?) that were exposed in some near-by land.

(5) Unstable bottom conditions, shown by slump structures, prevailed and may have been mainly the result of volcanic disturbances. Graywacke sediments and some graded bedding point to the activity of turbidity currents.

(6) No geographic limits to the depositional area and to the crustal depression on which the older complex was deposited are indicated by the Puerto Rican data. Woodring (1954, Fig. 1) suggests that in the Late Cretaceous almost the entire Caribbean region shared in the type of sedimentation exemplified by the Puerto Rican older complex and that volcanic vents were widespread over the entire region. Other writers prefer to limit the volcanic outpourings and deposition of geosynclinal sediments, as well as the ultimate orogenic deformation, to the narrow arcuate belt represented by the present island arc. (*See* Ewing and Worzel, 1954; Vening Meinesz, 1954; Hess and Maxwell, 1953.)

It is tempting to draw on analogy between the volcanic arc of the Lesser Antilles today and the Greater Antillean arc in older complex time. There is little reason to doubt a certain parallel. However, from signs of marine deposition interspersed throughout the section, it seems certain that the Puerto Rican older complex was deposited on a subsiding foundation. The essential structure would thus have been geosynclinal, whereas as far as can be discovered from surface indications and gravity data, the Lesser Antillean arc today is essentially geanticlinal.

(7) The older complex was deformed by intense faulting and relatively moderate folding sometime in early Tertiary time (probably the Eocene). The major fold axes trend generally northwestward. Except for an overthrust in the San Juan area, the writer recognized no good evidence of strong tangential stresses. The northwesterly regional strike may have been imposted by northeast-southwest-regional compression, such as would be generated in a tectogene with a northwest orientation (*see* Vening Meinesz, 1954), or by the slippage of the geosynclinal sediments into a deep northwesterly oriented grabenlike downwarp, perhaps like the present Puerto Rican Trench (Ewing and Heezen, 1955).

(8) The major plutons are differentiated complexes and seem to be roughly concordant with the strike of the country rock. This probably represents a squeezing aside of the country rock, although it may represent a selected accommodation by intrusion to pre-existing structures. The plutons therefore seem to be the youngest rocks in the older complex. The average composition of the plutons appears to be more silicic than the volcanic rocks of the older complex. If the former are derived from the latter, or if both are derived from the same

parent material, then either magmatic differentiation on a large scale has occurred, or there has been an addition (from where ?) of more silicic material.

(9) The width of the Antillean geosyncline and orogenic belt is probably not indicated by the Puerto Rican data. Some question exists, however, concerning the significance of the essentially east-trending structural alignment in the southwestern corner of the island. This structural deviation from the main northwest trend may be due to displacement by a large intrusive body lying off the south coast, or it may be the result of deformation outside the confines of the northwest-trending downwarp. The close association of serpentine (Pl. 1) with this zone is interesting in light of Hess' (1948) hypothesis that serpentine is generally intruded in two belts, 50 miles on either side of a tectogene axis, and that this intrusion takes place very early in the development of the downwarp.

(10) Data on major faults originating with the early Tertiary orogeny are unreliable owing to difficulty in distinguishing these from faults produced by the post-Paleocene tectonic disturbances. In the San Juan area, however, the relative absence of faults in the middle Tertiary rocks and their abundance in older complex rocks immediately adjoining (Pl. 1) points to an early Tertiary age for most of these structures. This seems to be particularly true for the large thrust and for faults displaced by it. It is not known whether this thrust is an overthrust or underthrust.

(11) The present east-trending structural axis of Puerto Rico dates from the Oligocene. The distribution and lithology of upper Oligocene coarse clastics suggests that a narrow though lofty strip of land, which roughly coincided with the present island axis, existed during their deposition.

(12) On the basis of the geophysical data from the north coast, the middle Tertiary deposits were probably mildly deformed by folding during their deposition. Post-early Miocene folding is apparent on the south coast, and an uparching of the entire island along the east-trending axis must have been closely associated with this.

(13) The pattern of inferred joints in the dense limestones of the north-coast middle Tertiary belt can possibly be explained as tension breaks formed by the foundering of the Aguadilla re-entrant, an arm of the Puerto Rican Trench, and by arching along the east-trending island axis. Joints due to the former deformation may date from middle Tertiary time, and those due to the latter from middle Tertiary to late Tertiary or Pleistocene time.

(14) Several faults of large displacement cut middle Tertiary rocks. Again, the age of these faults is uncertain, and the possibility that they date from the Pliocene and Pleistocene must be recognized.

(15) The present relief of the island was formed in Pliocene and possibly early Pleistocene time by the arching of a surface of low relief. Block faulting is thought to have contributed substantially to this uplift. Many relics of the scarps formed by this faulting still exist.

(16) A study of shore-line features (Kaye, in preparation) indicated that crustal movements such as have occurred in Puerto Rico since the late Pleistocene have caused negligible disleveling. This is shown by the advanced development of erosional and depositional features caused by the existing sea level and by a satisfactory correlation of later Quaternary marine levels with those recorded in other parts of the world.

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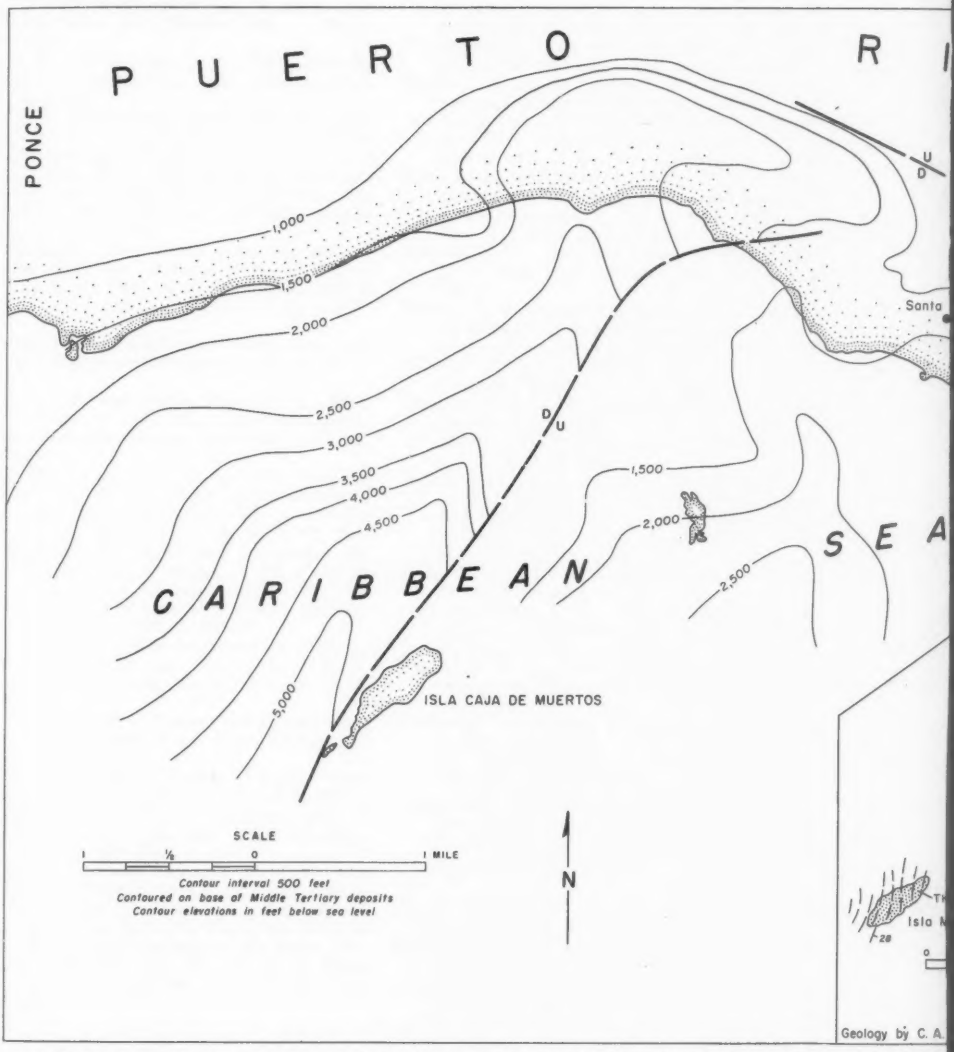
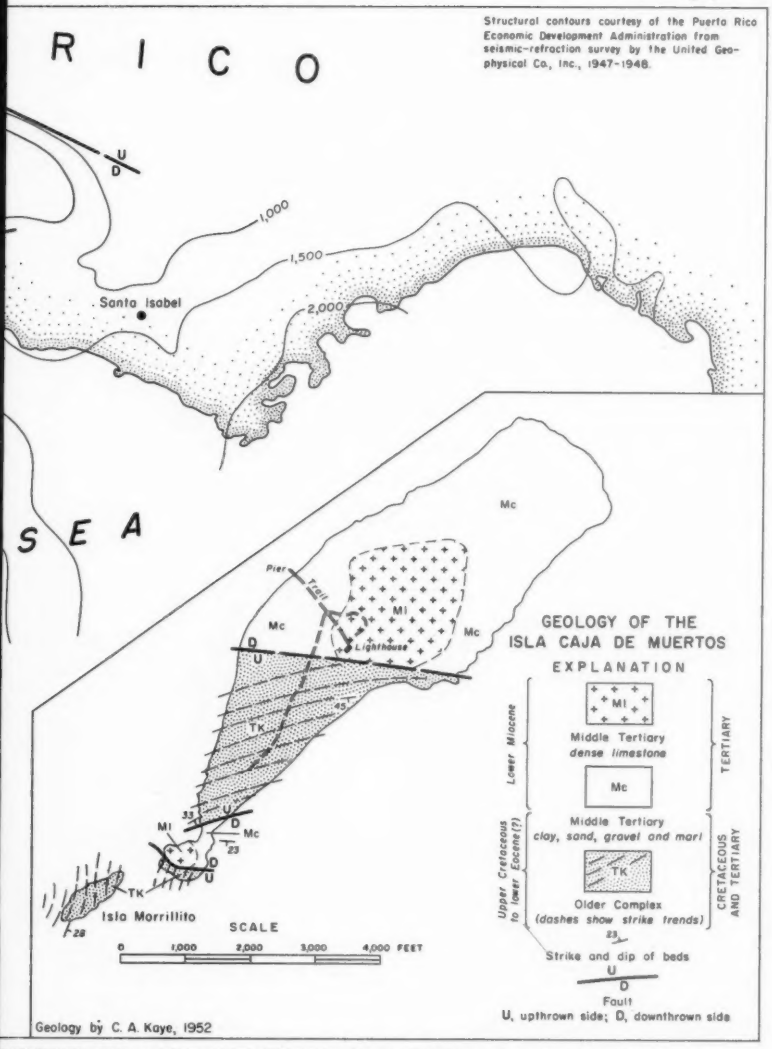


STRUCTURAL CONTOUR MAP OF PART OF THE SOUTH CO
REFRACTION SURVEY, AND GEOLOGY OF THE ISL



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Short Notes

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GRAVITY MEASUREMENTS IN ORIENTE PROVINCE, CUBA

BY G. LYNN SHURBET AND J. LAMAR WORZEL

In 1954 gravity measurements were made in Oriente Province, Cuba, at about 2-mile intervals from Guantanamo to Holguin via Santiago and Bayamo. Additional measurements were made between Guantanamo and Baracoa (Fig. 1.) These measurements supplement the 15 pendulum observations made in southern Oriente Province during 1933 and 1935 by the U. S. Coast and Geodetic Survey (Dickerson, 1940, p. 213).

A Frost geodetic gravimeter was used. The instrument has a range of about 4000 mgal for a single setting of the latitude adjustment. The reading spring is linear over the range used with a scale constant of 2.667 mgal per scale division. A reading is accurate to within 0.1 scale division, and drift is less than the reading error for periods up to several days.

Values of observed gravity were determined relative to a Frost gravimeter station at 18° 20.13′ N., 64° 57.29′ W., on St. Thomas, V. I., occupied 5 days after completion of the field work in Cuba. No direct connections with our laboratory base result from this work because of difficulties with the gravimeter during its transport from the laboratory to Cuba and from St. Thomas back to the laboratory. However, the St. Thomas station (978.6835 cm/sec^2) was determined in 1952 directly from the Lamont Geological Observatory Base #1 (980.2584 cm/sec^2) by Shurbet and Ewing (1956, p. 515). The LGO Base has been connected to the U. S. Coast & Geodetic Survey base (980.118 cm/sec^2) in the Department of Commerce Building, Washington, D. C. These connections are probably accurate to within 1 mgal.

Gravity values from various sources for the Guantanamo Bay station are listed in Table 1. Values from the S-21 (Vening Meinesz and Wright, 1930, p. 56) and the S-48 (Collins, 1933, p. 16) cruises have been changed by +5 and +6 mgal, respectively, to base them on the Commerce Building Base. The agreement of the gravimeter value with the pendulum values from the S-21 and CONGER cruises is not completely satisfactory, but barely within the accuracies of the observations. No explanation is offered for the lack of agreement of the gravimeter value with the pendulum value from the S-48 cruise.

During these measurements in Cuba, stations were reoccupied when practical to determine drift of the gravimeter. In Table 2 the gravity values for all stations, corrected for instrument drift, are given. A number of stations were occupied two or more times as indicated. Inspection of these values shows that a maximum error of about 0.8 mgal may be attributed to undetected instrument drift during this period. Drift during the 5-day interval between completion of the field work in Cuba and the occupation of the St. Thomas base station is considered negligible.

Station elevations in Cuba were determined by aneroid barometer with reference to four Inter-American Geodetic Survey bench marks (at stations 16, 29, 31, 36) and shoreline locations where elevations could be estimated from the sea surface. Repeated aneroid observations at various stations indicate that errors up to 50 feet may occur, corresponding to an error in simple Bouguer anomaly of about 3 mgal.

Positions for the stations were determined by speedometer distances from towns or river crossings, which could be located on the maps. The positions are probably accurate to within half a mile, corresponding to an error in gravity anomaly of about .5 mgal. The Inter-

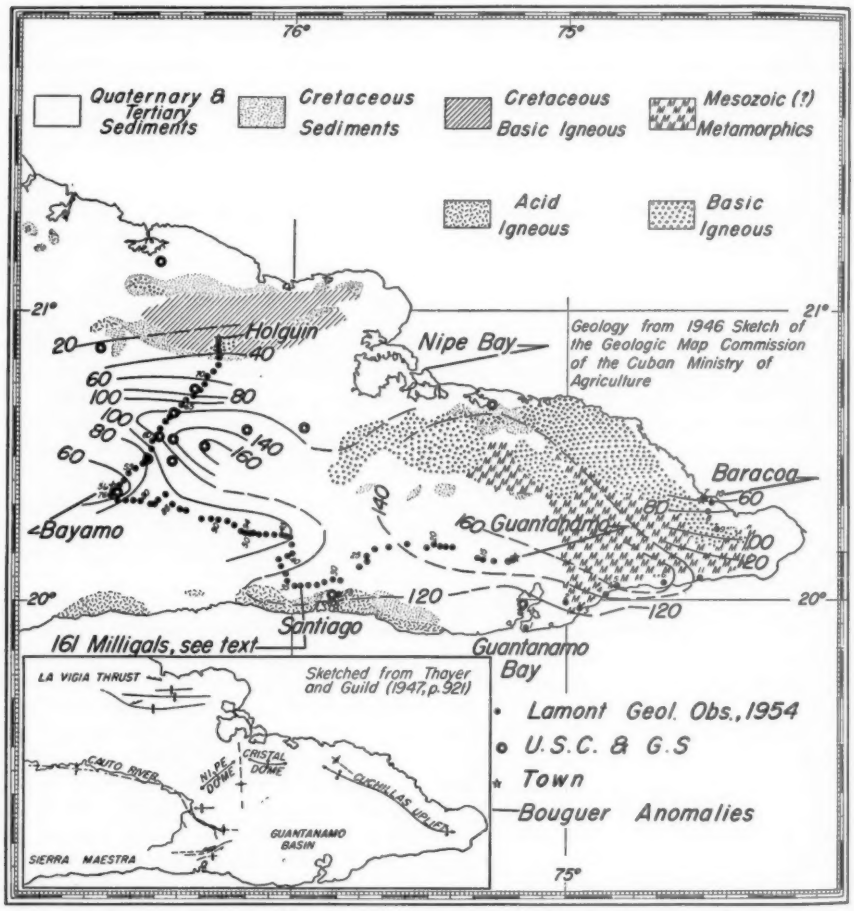

FIGURE 1.—GRAVITY MEASUREMENTS IN SOUTHERN ORIENTE PROVINCE, CUBA
Surface geologic formations adapted from a map by Brodermann, Albear, and Andreu.

TABLE 1.—OBSERVED GRAVITY VALUES AT GUANTANAMO BAY, CUBA

Observation	Date	Latitude	Longitude	Elevation (Meters)	Gravity
S-21*	1928	19° 54.5 N.	75° 08.9 W.	0	978.748
S-48**	1932	19° 54.5 N.	75° 08.9 W.	0	978.745
USS CONGER, Cruise III***	1954	19° 54.5 N.	75° 08.9 W.	0	978.748
Frost gravimeter	1954	19° 54.5 N.	75° 08.9 W.	1.5	978.7528

* Vening Meinesz and Wright (1930, p. 56).
** Collins (1933, p. 16).
*** Unpublished data, Lamont Geological Observatory.

TABLE 2.—PRINCIPAL FACTS FOR GRAVITY OBSERVATIONS
Base station Commerce Building Base, 980.118 cm/sec². Observer Hugh Traphagen.

STATION NUMBER	DATE 1954	LATITUDE NORTH	LONGITUDE WEST	ELEVATION FEET	OBSERVED GRAVITY GALS	INTERNATION FORMULA GRAVITY	FREEAIR ANOMALY IN MGALS	BOUGUER ANOM. MGALS
1	3 Feb				978.7531			
"	10 Feb				.7523			
"	"				.7528			
"	11 Feb				.7528			
1*		19-54.5	75-08.9	5	.7528	978.6464	107	107
2	3 Feb				978.7814			
"	4 Feb				.7816			
"	6 Feb				.7811			
2*		19-59.5	75-00.0	70	.7814	978.6512	137	134
3	4 Feb				.7827			
3	6 Feb				978.7822			
3*		19-58.1	74-57.0	5	.7825	978.6498	133	133
4	4 Feb	20-01.1	74-50.9	5	.8008	.6527	149	148
5	"				.8219			
"	6 Feb				.8219			
5*		20-02.5	74-49.0	30	978.8219	978.6541	171	170
6	4 Feb				.8267			
"	6 Feb				.8267			
6*		20-03.5	74-38.5	80	.8267	978.6551	179	176
7	4 Feb				.7262			
7	5 Feb				978.7264			
"	"				.7270			
"	6 Feb				.7275			
7*		20-21.1	74-29.7	15	.7268	978.6722	56	56
8	5 Feb	20-18.5	74-29.0	60	.7464	.6696	82	80
9	5 Feb				978.7510			
"	6 Feb				.7518			
9*		20-14.5	74-27.0	110	.7514	978.6657	96	92
10	5 Feb	20-20.0	74-28.5	15	.7358	.6711	66	66
11	6 Feb	20-04.5	74-30.5	5	.7920	.6560	137	136
12	6 Feb				978.8011			
"	7 Feb				.8016			
12*		20-08.0	75-12.0	100	.8014	978.6594	151	148
13	"				.8014			
"	9 Feb				.8014			
13*		20-08.0	75-12.5	110	978.8014	978.6594	152	148
14	7 Feb	20-08.0	75-16.0	160	.8051	.6594	161	155
15	"	20-08.5	75-18.0	175	.8038	.6599	160	154
16	"	20-08.5	75-19.5	209	.8019	.6599	162	154
17	"	20-11.0	75-25.0	500	.7800	.6623	165	147
18	7 Feb	20-11.0	75-26.5	480	978.7792	978.6623	162	145
19	"	20-11.0	75-28.0	480	.7779	.6623	161	143
20	"	20-11.5	75-29.0	465	.7790	.6628	160	143
21	"	20-11.0	75-30.5	550	.7734	.6623	163	143
22	"	20-11.5	75-35.0	555	.7736	.6628	163	143
23	7 Feb	20-11.0	75-39.0	800	978.7523	978.6623	165	136
24	"	20-10.5	75-42.0	895	.7432	.6618	166	133
25	"	20-09.0	75-43.5	655	.7592	.6604	160	137
26	"	20-08.0	75-45.5	565	.7592	.6594	153	132
27	"				.7510			

*Information adopted for this station

TABLE 2.—*Continued*

STATION NUMBER	DATE 1954	LATITUDE NORTH	LONGITUDE WEST	ELEVATION FEET	OBSERVED GRAVITY GALS	INTERNATION FORMULA GRAVITY	FREE AIR ANOMALY IN MGALS	BOUGUER ANOM. MGALS
27	9 Feb				978.7507			
27*		20-07.0	75-45.5	650	.7509	978.6584	154	130
28	7 Feb				.7891			
"	8 Feb				.7891			
28*		20-01.0	75-50.0	15	.7891	978.6526	138	137
29	8 Feb				978.7768			
"	9 Feb				.7768			
29*		20-04.0	75-49.5	213	.7768	978.6555	141	134
30	8 Feb	20-04.0	75-51.0	240	.7723	.6555	139	131
31	"	20-03.5	75-53.0	301	.7683	.6551	142	131
32	8 Feb	20-03.0	75-54.5	320	978.7688	978.6546	144	133
33	"	20-03.0	75-56.5	445	.7624	.6546	150	134
34	"	20-03.0	75-58.0	380	.7936	.6546	175	161
35	"	20-03.0	75-59.5	715	.7400	.6546	153	127
36	"	20-04.0	76-01.0	840	.7315	.6555	155	125
37	8 Feb	20-05.5	76-01.5	675	978.7403	978.6570	147	122
38	"	20-07.0	76-02.5	600	.7408	.6584	139	117
39	"	20-08.5	76-03.0	775	.7251	.6599	138	110
40	"	20-09.0	76-01.0	615	.7326	.6604	130	108
41	"	20-10.0	76-00.0	535	.7358	.6613	125	105
42	8 Feb	20-11.5	76-00.0	570	978.7328	978.6628	124	103
43	"	20-13.0	76-00.5	500	.7350	.6643	118	100
44	"	20-13.0	76-01.5	615	.7286	.6643	122	100
45	"	20-13.5	76-02.5	840	.7118	.6648	126	96
46	"	20-13.5	76-04.0	790	.7136	.6648	123	95
47	8 Feb	20-13.5	76-06.0	995	978.6992	978.6648	128	92
48	"	20-13.5	76-07.5	890	.7040	.6648	123	91
49	"				.7120			
"	9 Feb				.7123			
49*		20-14.0	76-08.5	750	.7122	978.6652	118	90
50	8 Feb	20-14.0	76-09.5	470	978.7248	978.6652	104	87
51	"				.7123			
"	9 Feb				.7123			
51*		20-23.0	76-39.5	160	.7123	978.6740	53	48
52	8 Feb	20-24.0	76-38.0	230	.7152	.6750	62	54
53	8 Feb	20-25.0	76-37.0	205	978.7230	978.6760	66	59
54	"	20-26.5	76-36.0	180	.7339	.6775	73	67
55	"	20-27.5	76-34.5	270	.7390	.6784	86	76
56	"	20-28.5	76-33.0	175	.7459	.6794	83	77
57	"	20-30.0	76-32.0	200	.7488	.6809	87	80
58	8 Feb	20-31.5	76-31.0	175	978.7696	978.6824	104	97
59	"	20-33.0	76-31.0	110	.7870	.6838	114	110
60	"	20-34.0	76-30.0	115	.8016	.6848	128	123
61	"	20-35.5	76-29.0	115	.8080	.6863	133	128
62	"	20-37.0	76-28.0	115	.8107	.6878	134	130
63	8 Feb	20-38.0	76-27.0	120	978.8096	978.6888	132	128
64	"	20-39.0	76-25.5	135	.8043	.6898	127	122
65	"	20-40.5	76-24.5	125	.7870	.6913	108	103
66	"	20-41.5	76-23.5	150	.7731	.6922	95	90
67	"				.7624			
67	9 Feb				978.7624			
67*		20-42.5	76-22.0	150	.7624	978.6932	83	78
68	8 Feb	20-43.5	76-20.5	195	.7534	.6942	78	70
69	"	20-45.0	76-19.5	205	.7467	.6957	70	63
70	9 Feb	20-46.5	76-19.0	270	.7382	.6972	66	57

* Information adopted for this station

TABLE 2.—*Continued*

STATION NUMBER	DATE 1954	LATITUDE NORTH	LONGITUDE WEST	ELEVATION FEET	OBSERVED GRAVITY GALS	INTERNATION FORMULA GRAVITY	FREE AIR ANOMALY IN MGALS	BOUGUER ANOM. MGALS
71	9 Feb	20-47.5	76-17.5	355	978.7246	978.6982	60	47
72	"	20-49.0	76-16.5	375	.7214	.6997	57	43
73	"	20-50.5	76-16.5	330	.7232	.7012	53	41
74	"	20-52.0	76-16.5	380	.7190	.7027	52	38
75	"				.7152			
75	9 Feb				978.7152			
75*		20-53.0	76-16.5	420	.7152	978.7037	51	36
76	"	20-22.0	76-40.0	135	.7155	.6730	55	50
77	"	20-21.0	76-38.5	150	.7283	.6721	70	65
78	"	20-21.0	76-37.0	190	.7451	.6721	91	84
79	9 Feb	20-21.0	76-35.0	205	978.7483	978.672	96	88
80	"	20-20.5	76-33.5	245	.7446	.6716	96	87
81	"	20-20.0	76-31.5	210	.7470	.6711	96	88
82	"	20-20.0	76-30.5	290	.7480	.6711	104	94
83	"	20-21.0	76-29.5	320	.7496	.6721	108	96
84	9 Feb	20-22.0	76-28.0	355	978.7510	978.6730	111	99
85	"	20-20.0	76-26.5	430	.7430	.6711	112	97
86	"	20-19.0	76-25.0	545	.7312	.6701	112	93
87	"	20-18.5	76-23.5	525	.7312	.6696	111	92
88	"	20-17.0	76-20.0	515	.7318	.6682	112	93
89	9 Feb	20-17.0	76-18.0	435	978.7323	978.6682	105	89
90	"	20-17.0	76-16.5	410	.7334	.6682	104	89
91	"	20-16.5	76-14.5	375	.7376	.6677	105	92
92	"	20-15.5	76-13.0	330	.7411	.6667	105	93
93	"	20-14.5	76-11.5	380	.7374	.6657	107	94
94	9 Feb	20-14.0	76-10.5	455	978.7304	978.6652	108	92

*Information adopted for this station

national Formula of Gravity of 1930 was used in computing the gravity anomalies.

Considering the errors mentioned above, the simple Bouguer anomalies listed in Table 2 are accurate to within about 3.5 mgal. Since a density of 2.84 was assumed in computing the Bouguer corrections (Worzel and Shurbet, 1955, p. 91), the gravimeter values are not strictly comparable to the anomalies from the pendulum observations for which a density of 2.67 was assumed. However, since the station elevations are small, the discrepancies are negligible.

A comparison of the simple Bouguer anomaly contours with a geologic map (Fig. 1), adapted from a map prepared in 1946 by Brodermann, Albear, and Andreu, shows no correlation of gravity anomalies with surface geologic formations. This lack of correlation may be explained by the insufficiency of gravity data everywhere except near the Cauto Trough, the area of the western gravity maximum. On the basis of a basaltic rock specimen collected about 6 miles east of the pendulum observation showing the largest gravity anomaly in this area, Dickerson (1940, p. 219) attributed this gravity maximum to the presence of an ultrabasic igneous mass lightly mantled with sediments. This indicated that the Cauto Valley is not a great syncline as formerly believed. The following structural and stratigraphic observations were made available to the authors by M. T. Kozáry, who is currently conducting field investigations of the structure of western Oriente Province, as evidence substantiating Dickerson's interpretation.

"1. It appears that the eastern half of the Cauto Plain was a positive structural area during the Upper Cretaceous and Eocene times since this area furnished detrital sediments into a northerly eugeosynclinal trough during the Upper Cretaceous and, in Eocene time, similar detrital sediments into the eugeosynclinal basin developing along the present site of the Sierra Maestra.

2. Although the Cauto Plain has been generally considered as a Tertiary syncline with its deepest portion located along the present site of the Cauto River, it can be shown that the plain must have

been essentially a stable platform area during Eocene time. The platform, shallow northward and with an abrupt southward facing ledge deepening into the Maestra trench, contained a relatively thin cover of Eocene sediments in the Cauto Plain. Part of the platform may have been eroded before the deposition of the Oligocene.

3. The type of Oligocene and Miocene sedimentation found in the southern rim of the Cauto Plain differs markedly from the sediments encountered along the north, the former representing essentially a molasse type detrital suite whereas the northern sediments are a shallow, calcareous and marly suite.

4. The center of the Cauto Plain contains, besides Neogene sediments in a general synclinal distribution, inliers of Paleocene sediments, found only 20 km south of the Cauto River, and Upper Cretaceous reefal limestone about 15 km south of the Cauto River."

Recent gravimetric work by private companies, also provided to the authors by Mr. Kozáry, shows that this area of high gravity anomaly ends abruptly to the west at about 20° 40′ N., 76° 40′ W. About 10 miles northwest of this point a much less pronounced gravity maximum of about 40 mgal is present.

The gravity anomaly of +161 mgal at station 34 (about 20° N., 76° W.) is about 30 mgal higher than anomalies only 1.5 miles both east and west of this station. This unusually large gravity gradient can be explained by an intrusive body not shown in Figure 1. Keijzer (1945, Dissertation, Royal Univ. Utrecht, p. 122) mentions "a thick bed or laccolith of basalt" very near this station which could account for the higher gravity anomaly, but, to account for the high gradient, the thickness of the body must be several times its diameter. Lewis and Straczek (1955), on detailed geologic maps of south-central Oriente Province show six small (largest dimension less than half a mile) intrusive bodies outcropping within about 5 miles of station 34.

We wish to express our appreciation to M. T. Kozáry for his contribution to this paper. Capt. G. M. Holly, Lt. S. T. Payne, H. H. Cole, R. E. Zaizer, and A. T. Jane at the U. S. Naval Base, Guantanamo Bay, and Manuel de Jesus Casalla, Capitan de Guardia Rurales, were of great assistance in arranging transportation, obtaining charts and equipment, and in providing local information to the gravity observer. H. Traphagen conducted the field work, and A. Trefzer made the drawing. This work was partially supported by the Office of Naval Research, Department of the Navy, under Contract N 6-onr-271 T. O. 8.

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MICRORELIEF OF THE SANTA MONICA SHELF, CALIFORNIA

BY RICHARD D. TERRY AND ROBERT E. STEVENSON

Introduction

In June 1955 the Allan Hancock Foundation, University of Southern California, began an oceanographic investigation to determine the feasibility of laying two sewer outfalls, 5–6 miles long, in Santa Monica Bay. A smooth sea floor with an even seaward gradient and no outcrops of rock or significant relief was desirable for the planned 12-foot-diameter concrete pipe. Rock along the line would require excavation in deep water, an expense to be avoided if possible. Information about the sediment texture was needed to design adequate foundations.

The bathymetry of Santa Monica Bay is based on U. S. Coast and Geodetic Survey smooth sheets, with additional soundings obtained by the VELERO IV, research vessel of the Hancock Foundation. The bathymetric chart (Fig. 1) was drawn, with a contour interval of 10 feet, to a depth of 300 feet. Although this interval does not permit micro-irregularities to be shown nor even an adequate presentation of "peaks" as high as 40 feet, the density of soundings does not permit use of a closer interval.

The major physiographic areas in Santa Monica Bay are: (1) the shelf, (2) Redondo and Santa Monica submarine canyons, (3) the basin slope, and (4) Santa Monica Basin.

This report concerns the central part of the shelf and, to a minor extent, the head of Santa Monica Canyon and the basin slope.

Microrelief

Generally, the surface of any continental shelf has small features consisting of mounds, ridges, depressions, or undulations called *microrelief*. Features with a relief of 3 feet are about the smallest that can be determined by an echo sounder because of the motion of the ship caused by sea and swell. Ripple marks and mounds formed by organisms are, therefore, usually too small to be recorded.

The VELERO IV made many traverses across the Santa Monica Shelf obtaining continuous echo soundings on a 300-foot recording instrument. Each fathogram was examined for various type of microrelief, and the character and types of relief were plotted along the track lines. From these data the general areas of smooth (Pl. 1, fig. 1), variable (Pl. 1, fig. 2), and irregular (Pl. 1, fig. 3) topography could be distinguished and their boundaries delineated. Track lines along which fathograms were obtained are shown in Figure 2.

More than 500 bottom samples were collected using snappers, coring tubes, scoopfish, jets, dredges, and divers, bottom photographs, and fathograms. The large number of bottom samples and hundreds of miles of sounding lines made it possible to correlate the bottom topography with the character of the bottom.

Types and Areas of Microrelief

Smooth zone.—A wide belt of smooth topography extends from near shore to varying distances out on the shelf (Fig. 3). The only distinguishable irregularities within this area are isolated hillocks a few feet high. Depressions are conspicuously absent within this zone. Small slope changes and flat areas (terraces) occur, but little microrelief is evident.

As might be expected, the depth increases gradually with increasing distance from shore. Near the edge of the shelf are occasional well-defined terraces, some of which can be seen on the bathymetric chart of Santa Monica Bay (Fig. 1). An especially well-defined terrace marks the boundary of the smooth and irregular zones and extends a short distance north toward the head of Santa Monica Canyon.

Quite possibly portions of the basin slope beyond the shelf and between the two submarine canyons are also relatively smooth, and a few fathograms indicate a smooth upper slope in this general area; however, since few fathograms extend below 270–300 feet, the nature of

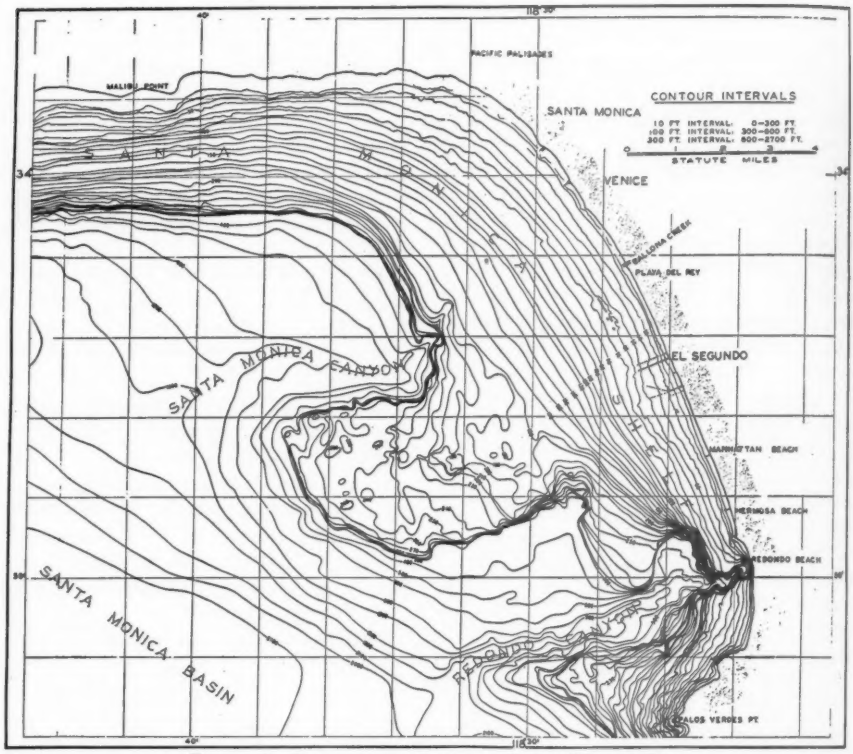

FIGURE 1.—BATHYMETRIC CHART OF SANTA MONICA BAY
The basin slope is located between Santa Monica and Redondo canyons.

the surface relief below that depth is uncertain. Sediments are mainly silty sand with small patches of gravel, neither of which form any microrelief.

Variable zone.—The zone of variable topography is generally seaward of the smooth area. The offshore limit could not be accurately determined because of the transitory nature of the boundary and the lack of fathograms in certain parts of this region. Within this zone are broad areas of no microrelief and smaller sections having low mounds, undulations, and terraces. Also included is the increased gradient at the edge of the shelf and the irregular topography in the two submarine canyons (Pl. 1, fig. 2). Along the edge of this shelf are numerous notches, steps, and small terraces; and closer to shore, north of Santa Monica Canyon, are several gullies 20–30 feet deep. Although there are many irregularities within this general area, none compares in size and number with those of the irregular zone. The sediments in

PLATE 1.—TYPICAL FATHOGRAMS SHOWING VARIOUS TYPES OF MICRORELIEF

Depths are in feet. The motion of ship causes the small "pips". Especially large swells are shown in the right third of Figure 3.

FIGURE 1.—Smooth zone across inner shelf. Note the absence of microrelief.

FIGURE 2.—Variable zone. A profile across the flat part of shelf. The shelf break is at the left.

FIGURE 3.—The irregular zone has very pronounced microrelief. The slope of the left side of the "peak" in right center is approximately 1° 10′. The relief of the right third of fathogram is part of the variable zone and shows a broad arch.

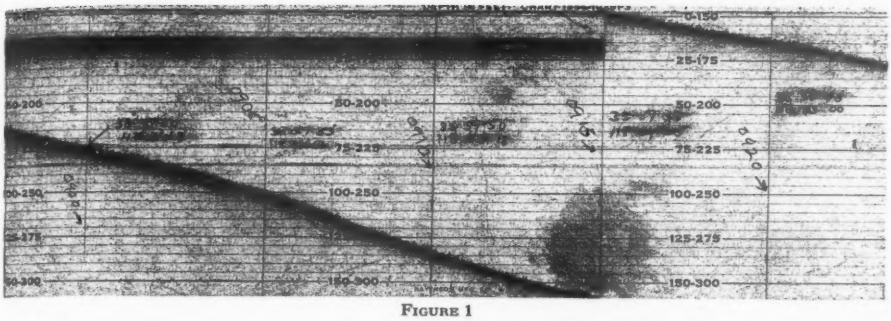

FIGURE 1

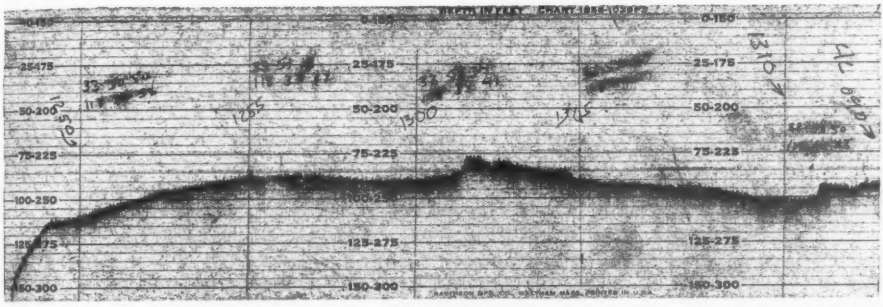

FIGURE 2

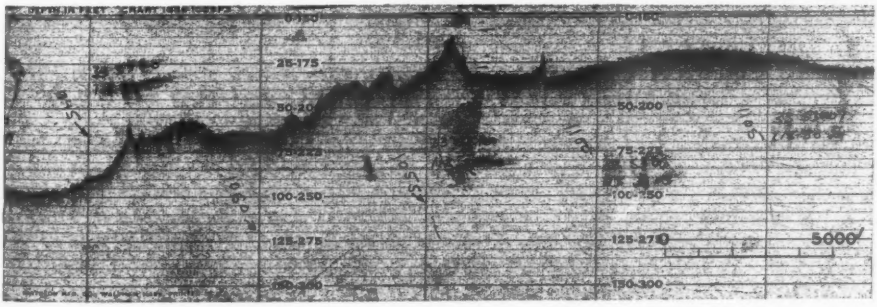

FIGURE 3

TYPICAL FATHOGRAMS SHOWING VARIOUS TYPES OF MICRORELIEF

this
irreg
is ev
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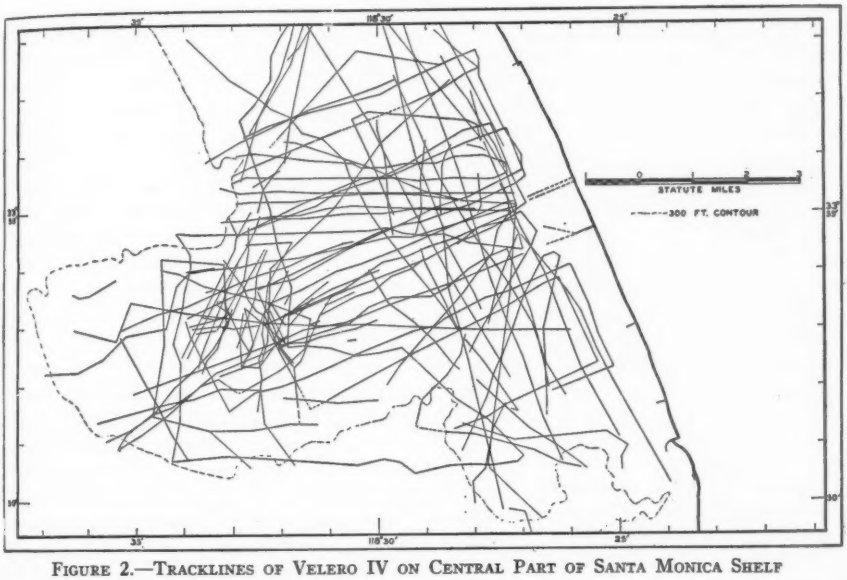

FIGURE 2.—TRACKLINES OF VELERO IV ON CENTRAL PART OF SANTA MONICA SHELF

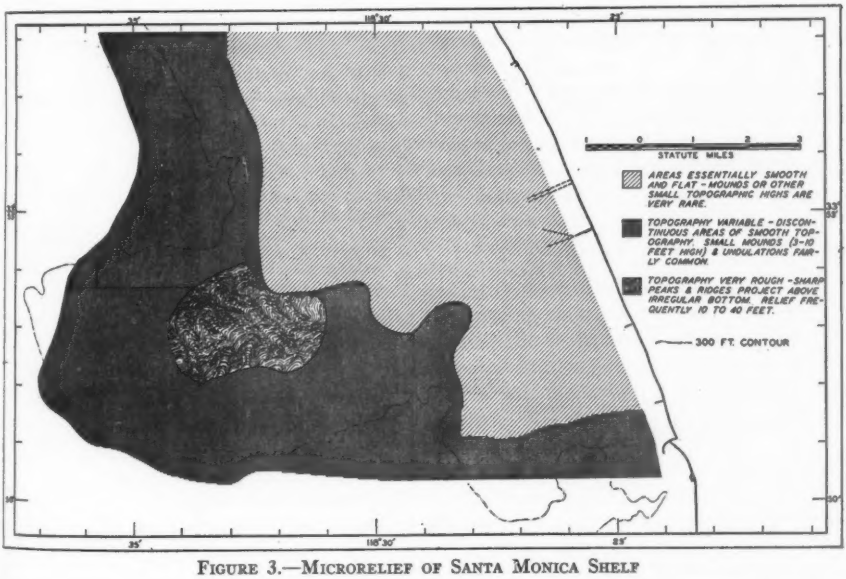
STATUTE MILES
AREAS ESSENTIALLY SMOOTH AND FLAT - MOUNDS OR OTHER SMALL TOPOGRAPHIC HIGHS ARE VERY RARE.
TOPOGRAPHY VARIABLE - DISCONTINUOUS AREAS OF SMOOTH TOPOGRAPHY, SMALL MOUNDS (5-10 FEET HIGH) & UNDULATIONS FAIRLY COMMON.
TOPOGRAPHY VERY ROUGH - SHARP PEAKS & RIDGES PROJECT ABOVE IRREGULAR BOTTOM. RELIEF FREQUENTLY 10 TO 40 FEET.
300 FT. CONTOUR

FIGURE 3.—MICRORELIEF OF SANTA MONICA SHELF

this zone vary from clays to gravel and are irregularly distributed. Some of the microrelief is evidently due to this irregular horizontal distribution.

Irregular zone.—The irregular zone consists mainly of mounds and ridges which project 10–40 feet and in places 80 feet above the surrounding sea floor. As noted in Figure 3 of Plate 1 the flanks of the mounds and ridges appear steep on fathograms. The slopes may be

more than 15°, but most are less than 1°–2°. Some of the mounds are connected by low ridges, and these also have irregular and jagged sloping sides. Individual mounds average 2000–3000 feet in diameter.

Bedrock as well as considerable amounts of rounded pebbles and cobbles have been dredged from this zone. Some of the mounds are isolated and surrounded by cobbles and gravel at their bases. The gravel is well rounded, and its areal extent is far greater than that of the projecting mounds. Most of the gravel is igneous, with a few widely scattered siliceous shale and metamorphic cobbles interspersed in the sediment.

Interpretation

The zone of smooth topography appears to be a depositional plain having material ranging in size from silt to gravel, but no rock crops out. Cores, jettings, and borings within the area indicate that the sediments vary widely in lithology in a vertical direction. Four deep jet holes on the shelf and numerous well holes on land near the shore show that sediments are as much as 500 feet thick in this zone. From these data it is believed that bedrock is rare or absent at or near the surface in the area of no microrelief. Nevertheless, rock may be near the surface at the outer periphery of this zone.

Even though the zone of variable topography also receives sediments, deposits do not cover all the irregularities formed when sea level was lower. The lack of "peaks" indicates either that sediments have at least partly covered many projections and that only remnants are left, or the rocks were more easily eroded than those in the irregular zone. Old stream channels across the shelf, notches, and gullies near the edge of the shelf probably were formed during a lower stand of sea level and have remained more or less unaltered. It is quite likely that rock exists in many places within this area, both at and near the surface.

The zone of rough and irregular topography is composed largely of rock and well-rounded gravel. A variety of bedrock has been dredged including shale, mudstone, and schist. Studies of the currents in Santa Monica Bay, including numerous measurements near the bottom, show that velocities of less than 0.2 knots are the most common. These currents should produce little erosion; therefore, the relief probably is due to differential erosion during a lower stand of sea level, the remnants having survived sedimentation and minor erosion during Recent times. Well-rounded gravel is probably a late Pleistocene residual deposit, and the presence of phosphorite, glauconite, and high calcium carbonate on the outer part of the shelf suggests slow sedimentation.

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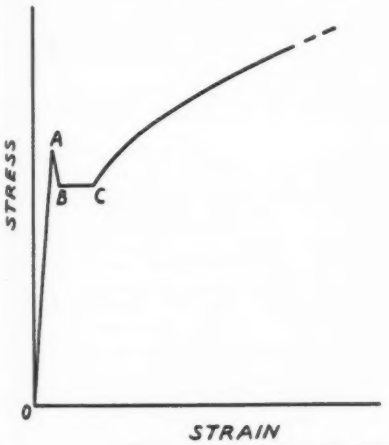
LÜDERS' BANDS AND PLASTIC DEFORMATION IN THE EARTH'S CRUST

BY M. S. PATERSON

Menard (1955) has suggested that the "fracture zones" in the Northeastern Pacific Basin demarcate a strain pattern in the plastically deformed crust which may be compared to the Lüders' bands that appear on a mild-steel test piece when it yields. If the deformation in the crust is, indeed, of the character of the Lüders deformation in mild steel, special stress-strain properties of the crustal material are implied.



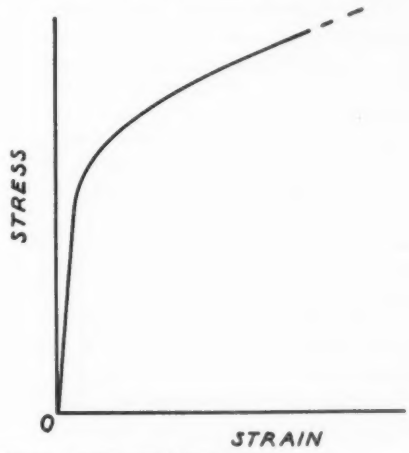


FIGURE 1 (left).—TENSILE STRESS-STRAIN CURVE FOR MILD STEEL
FIGURE 2 (right).—TENSILE STRESS-STRAIN CURVE FOR MATERIALS NOT HAVING A SHARP YIELD POINT

The peculiar behavior of mild steel and certain other materials (Cottrell, 1953), in which the distribution of plastic strain after yielding under a uniform stress is nonuniform, results from a stress-strain property best illustrated by their tensile stress-strain curves. In the region ABC (Fig. 1) the stress needed to continue plastic straining is lower than that needed to start it. This type of instability results in the concentration of the plastic deformation at first in those zones (Lüders' "lines" or bands) in which it began. Within the Lüders' bands, plastic deformation continues until local strain hardening raises the yield stress beyond the applied stress. Then yielding occurs in other, previously undeformed zones, until finally the whole specimen has yielded. Further deformation (beyond point C in Fig. 1) generally takes place uniformly throughout the specimen. The spacing of the Lüders' bands is often observed to be regular, but the reason for this is not properly understood; in some cases, for example in a thin wire, a single Lüders' band may appear at one end and gradually spread over the whole length of the specimen (Sylwestrowicz and Hall, 1951).

Most crystalline substance (for example copper, sodium chloride, limestone, and iron containing no carbon and nitrogen) have stress-strain curves like that illustrated in Figure 2. There is no drop in stress when yielding occurs, and the plastic straining occurs uniformly in a specimen under uniform stress. No Lüders' bands appear since any local concentration of plastic deformation would be immediately accompanied by additional strain hardening

which would raise the local yield stress above that necessary to continue plastic deformation elsewhere in the specimen.

Thus, if the "fracture zones" in the Northeastern Pacific Basin are Lüders' bands, the crustal material may be expected to have the special property that, when yielding begins at any given place, the local yield stress there falls immediately, as shown in Figure 1. Furthermore, since the Lüders' bands are the regions in which this local yielding occurs, the regions between the "fracture zones" cannot yet have undergone plastic deformation.

An explanation for the appearance of Lüders' bands at 45° to the directions of the principal stresses has been given by Taylor (1934), although in practice deviations from exactly 45° are commonly observed. The Lüders' bands are thus parallel (or nearly parallel) to the directions of maximum shear stress, the "slip lines" (or "shear lines") of the mathematical theory of plasticity, and they can be useful in delineating the plastic slip-line field (Nadai, 1950). However, Lüders' bands should not be identified with the mathematical slip lines, as is sometimes done (Scheidegger, 1953), since the latter are families of theoretical curves associated with a given stress pattern, and they are not delineated on the surfaces of most materials after straining. Further, the slip lines of the mathematical theory of plasticity should not be confused with the crystallographic slip lines of crystal plasticity; these are the intersections of glide planes with the surface, and they may or may not be parallel to the mathematical slip lines.

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